

TECTONIC HISTORY OF THE SOUTH-WEST PACIFIC REGION

by

JOHN RICHARD GRIFFITHS, M.A.

(University of Cambridge)

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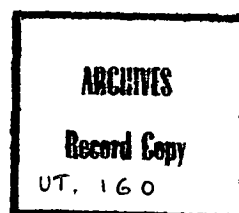
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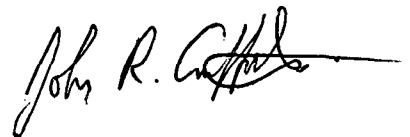
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(vol. 1)

This thesis contains no material which has been accepted for the award of any other degree or diploma in any university and to the best of my knowledge and belief contains no copy or paraphrase of any material previously published or written by another person except where due reference is made in the text of the thesis.

A handwritten signature in black ink, reading "John R. Griffiths". The signature is written in a cursive style with a long horizontal flourish extending to the right.

JOHN RICHARD GRIFFITHS
University of Tasmania,
Hobart.
June, 1973

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ABSTRACT

This study attempts to deduce the Phanerozoic tectonic relationships of Australia, Antarctica and New Zealand, and then to discuss in greater detail the evolution of New Zealand and of the continental margin of south-east Australia. Plate tectonics provides a broad conceptual framework for the syntheses proposed, but the emphasis throughout is based on objective analysis of the data available. Flow-charts and time-space plots are developed as aids in construction and evaluation of the tectonic syntheses.

Various criteria are used to define and reassemble Australia, Antarctica, New Zealand and surrounding areas of continental crust as parts of the former Gondwanaland in the Mesozoic. A brief review of the Paleozoic (exposed mainly in eastern Australia), outlines the progressive development and cratonisation of a very complex orogenic belt bordering the Pacific Ocean. During the Mesozoic an "Atlantic-type" continental margin was established east of this belt, and is now best exposed in New Zealand. Orogenesis along this margin accompanied subduction of the Pacific plate during the Upper Jurassic to Lower Cretaceous Rangitata Orogeny. The resulting orthotectonic belt exhibits contrasting structural and metamorphic styles of foreland and oceanward sides.

Fragmentation of Gondwanaland began in the Jurassic by development of an extensive rift valley system, which is studied in detail in the sedimentary basins of south-east Australia (Otway, Bass and Gippsland Basins), largely from sub-surface oil-company exploration data.

The structural evolution of these basins is controlled by the initial rift configuration, and neither pre-existing basement structures, nor transform faults related to Tertiary sea-floor spreading, appear to have much influence.

Published marine magnetic data constrain a geometric synthesis of the sea-floor spreading history. From 85 to 60 million years ago, the Tasman Sea was opening between Australia, Antarctica and New Zealand, about a rotation centre near northern Queensland. Following separation of Australia and Antarctica about 55 million years ago, a new three-plate geometry was established. In this system the Australian-Pacific plate boundary displays a complex history involving regions of subduction and transcurrent faulting. The Alpine Fault (which became active about 20 million years ago) and the severe Tertiary deformation in New Zealand are closely related to the sea-floor spreading history.

The various aspects of the regional tectonic history studied all indicate that the plate tectonic hypothesis is both a valid and useful new approach in geology. They are also predictive, and point to many new problems which might be investigated.

CHAPTER 1

GENERAL INTRODUCTION

This thesis aims to present an interpretation and synthesis of the tectonic evolution of the south-west Pacific region. The area studied comprises about one-eighth of the earth's surface, and includes eastern Australia, New Zealand, parts of Antarctica, and the surrounding oceans. The study concentrates mainly on the Mesozoic and Cenozoic history, but also includes some discussion of the earlier geology. The subject is discussed in three broad divisions, being respectively a regional study (Chapter 2), and detailed studies of New Zealand (Chapter 3) and south-east Australia (Chapter 4). A consideration of the general methods and principles of tectonics also forms an important part of the work. This aspect is treated briefly in the following Section, and a more comprehensive discussion is included as Appendix A.

1.1: TECTONICS AND GEOLOGY

Tectonics has been defined as the "science of the structure of the earth's crust, and the movements and forces which have produced it" (Dennis, 1967, p. 153). Badgely (1965, p. 1) distinguishes tectonics (geotectonics) from structural geology on the basis of scale - structural geology being "the study of individual structures - such as anticlines, thrust faults, lineation, and so forth - within a tectonic unit", and tectonics being "the study of the form, pattern and evolution of large-scale tectonic units [of the earth's crust] such as basins, disturbed belts, forelands, and continental shelves". Belousov (1962, p. v) states in regard to geotectonics that "this branch of geology has a somewhat peculiar status and

provides, in some measure, the theoretical basis of geology". Many other authors have also discussed tectonics and its relation to geology. Some attempt a rigid definition of tectonics, others see tectonics as a broader, more general, aspect of geology. Goguel (1962, p. 2) states "as a characteristic of geology in general, tectonics is principally a historical science, inasmuch as its object is the reconstruction of past phenomena". I therefore see little real difference in the aims of tectonics and geology, in that the study of both is ultimately aimed at understanding how an area has evolved. It is this context in which I will discuss the tectonics of the south-west Pacific region - the aim is to present a synthesis of the geological history. This approach to tectonics thus demands a study of all the geological data from the region, and development of a synthesis which should be compatible with all these data.

The advent of the plate-tectonic hypothesis over the last few years has led to a broad, conceptual framework within which such studies can be made. However, it is important to consider the theoretical basis of the approach to tectonic synthesis, so that the relationships between fact and theory are clearly understood. A fuller discussion of some of these theoretical aspects is presented in Appendix A, which gathers together some of the abstract ideas developed whilst preparing this thesis. It is not essential to read this Appendix first, but the concepts embodied in it have weighed heavily in the organisation and presentation of geological data and syntheses in the following Chapters.

In the light of these theoretical considerations, I have therefore tried throughout to apply an *objective* approach in constructing

tectonic syntheses. To do this, a basic methodology must be first established. The time-factor involved in geological processes virtually precludes experimental testing of tectonic syntheses, and thus it is especially important to use a logical approach, and to devise independent techniques for evaluation of proposed syntheses. A tectonic SYNTHESIS is the sum of ANALYSIS plus HYPOTHESIS. Analysis is the data collection and handling stage, which should be as *objective* as possible. In order to produce a synthesis, it is necessary to invoke or devise a hypothesis, and then to construct a model which simulates the geological data. Finally the synthesis should be rigorously *evaluated* against the original data. By careful separation of analysis and hypothesis in this way, the resulting synthesis will also be *predictive*, and point to new data which can be independently obtained to further evaluate the synthesis.

In this thesis I have selected "plate tectonics" as the broad, conceptual hypothesis within which models will be constructed, and then each is evaluated against various sets of data. In each Chapter, the basic layout is firstly to review the available geological (including geophysical) data, and then to develop and present a synthesis based on the plate-tectonic approach. Many predictions are explicit or implicit in the text and figures, and some degree of evaluation is also attempted. However, the final assessment of the validity or otherwise of the syntheses presented must be made independently. The predictions are perhaps the real conclusions of this thesis, in that they provide a means of evaluating the ideas expressed here, and provide a basis for planning of future investigations.

1.2: SCOPE, PRESENTATION AND LAYOUT

The area studied is conveniently referred to as the south-west Pacific, and includes eastern Australia, New Zealand and that part of Antarctica to the south, together with the south-east Indian Ocean, Tasman Sea and western South Pacific Ocean. The main part of the work concerns the Mesozoic and Cenozoic geological history of these areas, with some references to the earlier history and to peripheral regions. The study originally began as a review and synthesis of the evolution of the continental margin of south-east Australia, in the context of continental drift. This involved establishing pre-drift positions of surrounding continental fragments, and consideration of the drift history. This led to construction of a regional sea-floor spreading model, also involving the New Zealand region. An earlier interest in the Mesozoic geology of New Zealand was also continued, and this was later extended to a review of the New Zealand Cenozoic in the light of the sea-floor spreading model. The study is centred very largely on published literature, and also much unpublished data for south-east Australia. It has been necessary to read, review and interpret much work, and to use these data as the building-blocks for the tectonic syntheses, and at the same time to explore the conceptual possibilities of plate tectonics. The final result is an integration of these two lines of work.

In order to preserve continuity in the text, the order of actual work has been disregarded in favour of a more logical approach. Thus the regional evolution is dealt with first (Chapter 2), and then two separate themes developed - New Zealand (Chapter 3) and South-East Australia (Chapter 4) - which only indirectly relate to each other. Most of the

basic data, details of working methods, and the syntheses are contained in the figures, and the text is really an extended commentary based on the figures, giving opportunity for discussion of various points in more detail.

Pages in the text are numbered separately for each Chapter in the *top* right-hand corner. The many cross-references necessary are more conveniently made to Sections of Chapters, rather than pages and hence each page also has a number in the *bottom* right-hand corner which indicates the Section beginning or current on each page, to facilitate cross-reference. Tables are inserted in the text. Figures are again numbered separately, in the top right-hand corner, for each Chapter. References are cumulative for text and figures (see also notes on pages 4.2 and 4.3 regarding unpublished work in Chapter 4).

1.3: ACKNOWLEDGEMENTS

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Esso Australia Ltd. provided copies of a great deal of their unpublished data, which is greatly acknowledged. Copies of unpublished work have also been given by Mr. P. Bollen (Woodside-Burmah Oil N.L.); Mr. R.B. Leslie (Interstate Oil Ltd.); and Mr. B. Stinnear (Bureau of Mineral Resources) who also provided assistance in searching through unpublished oil exploration reports. I also thank the staffs of Esso Australia Ltd. (South-east Coastal Division), the Geological Surveys of New South Wales, South Australia, Tasmania and Victoria, the Bureau of Mineral Resources, various Australian Universities, and also many other individuals, with whom I have had the opportunity to discuss aspects of this work.

Mrs. D. Stuetzel compiled the bathymetry for Fig. 2.1. I thank the Photographic Section, University of Tasmania, for their assistance. Typing facilities were made available in the Geology Department, University of Tasmania, and the typing has been done by Miss J. Griffiths and Miss B. Albury, whom I thank.

CHAPTER 2

REGIONAL TECTONICS OF THE SOUTH-WEST PACIFIC

In the context of this thesis the south-west Pacific region is taken as the area roughly enclosed by the 120°E and 140°W meridians, the equator, and the South Pole. It includes Australia, New Zealand and part of Antarctica as the major land masses, and the Tasman Sea and parts of the Indian and South Pacific Oceans as the major ocean basins (Fig. 2.1). Various lines of evidence (see below) indicate that much of the ocean floor is considerably younger than the continental areas, in accord with the general concept of continental drift. In this chapter I shall attempt to reconstruct the former positions of the continental areas prior to the most recent episode of drift, and then to interpret and synthesise the major geological features of the region in the light of these reconstructions.

The layout of this chapter, and the relationship of succeeding ones to it, is summarised by a flow-chart (Fig. 2.2), which emphasises the fundamental importance of the pre-drift Gondwanaland reconstruction in this thesis. If the basic assumption that continental drift has occurred is valid, it is necessary to establish accurately the pre-drift positions of the continental fragments involved (Section 2.2), working from data which are currently available (Section 2.1). This reconstruction then provides the framework within which any events which occurred prior to drift must be considered (Sections 2.3, 2.4). The chronology and nature of

subsequent movements can be established by geometrical analysis based on marine magnetic anomalies (Section 2.5), and this model used as a basis for study of other plate margins in the region (Section 2.6).

In this chapter I first present the evidence leading to the Mesozoic reconstruction of the Gondwanaland margin, and then outline the regional tectonic evolution in chronological sequence. Chapters 3 and 4 are concerned with more detailed analyses of New Zealand and south-east Australia. At this stage I will draw on the conclusions of these chapters as necessary to build up the regional synthesis, leaving detailed discussion till later. Much of the basic data, and many of the results of this regional study are most conveniently presented as maps and diagrams, with accompanying legends. The text is thus complimentary to the figures, allowing discussion of various aspects in more detail. Most of the maps are based on a Lambert's Equal-area Azimuthal projection, chosen to give a realistic visual picture with minimum distortion.

2.1: PRESENT TECTONIC FRAMEWORK OF THE SOUTH-WEST PACIFIC

The bathymetry of the south-west Pacific region is outlined in Fig. 2.1. Five morphologically different areas can be distinguished:-

- (a) The continental land masses of Australia (including New Guinea) and Antarctica, which are largely surrounded by relatively narrow continental shelves;

- (b) the deep oceanic areas of the Indian and South Pacific Oceans and the Tasman Sea;
- (c) the broad median ridge separating Australia and the Campbell Plateau from Antarctica;
- (d) the large, shallowly submerged areas north and east of the Tasman Sea, forming the Lord Howe Rise, Norfolk Ridge, Campbell Plateau, Chatham Rise and several smaller ridges, and emergent in New Zealand, New Caledonia, eastern New Guinea, the Solomon Islands and several other small islands;
- (e) the ridges and trenches of the Tonga, Kermadec, New Hebrides and Macquarie systems.

Geophysically it is possible to distinguish two basic types of crust - continental (about 25 - 50 km thick) and oceanic (about 5 km thick). Australia and Antarctica (a) are continental crust, and the deep ocean basins (b) are floored by oceanic crust (Officer, 1955). The shallow areas (d) north and east of the Tasman Sea are continental crust ranging up to about 30 km thick (Officer, 1955; Thomson & Evison, 1962; Solomon & Biehler, 1969; Shor *et al.*, 1971; Woodward & Hunt, 1971), and hence must be taken into account in any reassembly of continental fragments (see Section 2.2). The ridge-trench systems (e) north of New Zealand are broadly similar to other *island arcs* of the western Pacific, which have probably formed as a result of interaction of converging plates (Sykes, 1966; Oliver & Isacks, 1967; Isacks, Oliver & Sykes, 1968), and are not associated with any continental crust. The Macquarie Ridge is a more complex feature, but still results from oceanic plate interaction.

In terms of plate tectonics, the south-west Pacific region currently includes parts of three major plates - the Australian, Antarctic and Pacific plates (Fig. 2.3). [Le Pichon (1968) originally defined these as the *Indian*, Antarctic and Pacific plates. Australia is part of the Indian plate at the present day, but this has not always been so (c.f: McKenzie & Sclater, 1971). Hence to avoid confusion I will consider the Australian continent to be part of the *Australian* plate in this thesis]. The seismicity of the region (Fig. 2.3, based on Barazangi & Dorman, 1969) helps to delineate the present boundaries between these plates. The Australian/Antarctic and Pacific/Antarctic boundaries are mid-ocean ridges (Fig. 2.1), associated with shallow seismicity (Fig. 2.3), at which active sea-floor spreading is taking place (Le Pichon, 1968). The Australian/Pacific plate boundary runs along the Macquarie Ridge, through New Zealand and along the Kermadec/Tonga Trench (Figs. 2.1, 2.3). The southern part of this boundary is complex and variable (Hayes *et al.*, 1972; Hayes & Talwani, 1972), but north of New Zealand it is a major subduction zone at which the Pacific plate is being consumed (Oliver & Isacks, 1967; Le Pichon, 1968). The boundary then swings west towards New Guinea, passing into a complex region of many small plates, and finally linking up with the Eurasian/Pacific plate boundary in this zone.

Marine magnetic profiler results are available across the Pacific/Antarctic and south-east Indian ridges (Le Pichon & Hiertzler, 1968; Pitman *et al.*, 1968; Christoffel & Ross, 1970; Herron, 1971; Weissel & Hayes, 1971, 1972; Falconer & Christoffel, 1972), and from the Tasman Sea (Hayes & Ringis, in press). These

data are summarised in Fig. 2.4. The magnetic results show that the oceanic crust between Australia and Antarctica is 55 to 0 MY old (Weissel & Hayes, 1971); between the Campbell Plateau and Antarctica 83 to 0 MY old (Christoffel & Falconer, 1972); and in the central Tasman Sea 77 to 60 MY old (Hayes & Ringis, in press). [The abbreviation MY is used throughout this thesis for "million years"].

Le Pichon (1968) attempted an analysis of the vectors of relative motion across plate boundaries in the south-west Pacific and elsewhere, based on the magnetic profiler data available at the time. With an increased amount of data available, new determinations of the relative vectors and instantaneous poles of rotation between the south-west Pacific plates have recently been made (e.g: Christoffel, 1971; McKenzie & Sclater, 1971; Christoffel & Falconer, 1972; Hayes & Talwani, 1972; Weissel & Hayes, 1972). The general trends of the vectors of motion across the plate margins are indicated in Fig. 2.3. Only two models are shown, but all of the others are essentially similar (c.f: Hayes & Talwani, 1972, fig. 7).

Geological data is available from all the areas not covered by sea or ice. The major geological elements of the region are shown in Fig. 2.5, and will be discussed in more detail later. In addition, recent results from the Deep Sea Drilling Project (1972) indicate minimum ages for the ocean floor at several sites, also shown in Fig. 2.5.

The preceding brief summary outlines the general nature and extent of geological and geophysical data currently available

from the south-west Pacific region. Although there is still much more basic information to be collected, the data currently available are sufficient to delineate the broad tectonic framework of the region, to provide several constraints on any attempt at continental reconstructions and to delineate areas from which more data are most needed.

2.2: RECONSTRUCTION OF THE SOUTH-WEST PACIFIC MARGIN OF GONDWANALAND

Early reconstructions of Gondwanaland were based mainly on rough morphological matching of the outlines of the major continents—Antarctica, South America, Africa, India and Australia, and the recognition that when so joined together many major geological features appeared to continue from one continent to another, seemingly irrespective of the oceans which now separate them (e.g. Wegener, 1924; Du Toit, 1937). Carey (1958), using a more rigorous approach, suggested that the morphological fits were more correctly made using the edge of the continental crust, rather than the present shorelines, and thus he obtained much better fits by using an intermediate submarine contour, generally 2000 metres. Bullard *et al.* (1965) used the 500 fathom contour to match up the continents around the Atlantic Ocean, using a computer to determine the positions of "best fit". Sproll & Dietz (1969) performed a similar analysis, using the 1000 fathom contour, for the Australia/Antarctica fit, and Smith & Hallam (1970) presented a computer fit of all the major components of Gondwanaland at the 500 fathom contour. None of these fits have however paid much attention to the New Zealand

region, except generally to place it closer to Australia.

New Zealand is cut in two by a major transcurrent fault (the Alpine Fault) across which there is a recorded dextral offset of 480 km (Wellman, *in* Benson, 1952). Geological reconstructions of New Zealand by Carey (1958), Quennell (1964) and Fleming (1970) have reversed this displacement and attempted to restore New Zealand to its probable Mesozoic configuration. Attempts have then been made to integrate the reconstructed form of New Zealand with the Australia/Antarctica fit (e.g: Carey, 1958; Wright, 1966; Van der Linden, 1967, 1969; Summerhayes, 1969; Cullen, 1970; Jones, 1971; Suggate, 1972). These reconstructions in general fail to fully utilise and in some cases are inconsistent with data obtained from bathymetric, geophysical and geological studies, and are therefore an inadequate basis for detailed tectonic discussions. I have previously attempted (Griffiths, 1971a,b) to morphologically reconstruct the continental fragments of Gondwanaland now distributed in the south-west Pacific in a way which is consistent with all the available data, and will now further discuss and analyse this reconstruction.

2.2.1: *BASIC PRINCIPLES OF CONTINENTAL REASSEMBLY*

The reassembly of continental fragments which are believed to have been once joined together has essentially three stages:

- (a) definition of the fragments to be reassembled
- (b) morphological fit of the fragments (if necessary reversing any post-drift deformation);
- (c) test of the geological and geophysical credibility of the proposed fit.

The actual fit (b) need not be obtained by pure "trial and error", but can utilise available information from marine geophysics, land geology, etc, as a guide, thus satisfying part of (c) during the process.

The basic inter-relationships of the parameters involved in continental reassembly are illustrated in the form of a flow-chart (Fig. 2.6) for the general case. The area enclosed by the dashed line comprises the basic module for drift reconstruction. *Input* into this module consists of bathymetric, geophysical and geological data; *output* is an acceptable reconstruction of the fragments (assuming that this is possible). The upper half of the module concerns only the definition of the fragments. The basic assumption which allows reassembly to take place is that the sea-floor spreading hypothesis is correct. The lower part of the module concerns the actual operation of reassembly, and the application of geological and geophysical tests of its credibility. It may be necessary to go around this part of the circuit several times in order to achieve an acceptable fit, consistent with all the data fed into the module. If no such fit is obtained, then the input data should be checked carefully for accuracy. If a fit is still unobtainable, then it may be that either (a) the sea floor spreading hypothesis is wrong, or (b) the fragments never fitted together, even though sea-floor spreading has occurred.

This basic scheme for continental reassembly can be used both to develop new reconstructions, and to critically evaluate any reconstructions based upon intuitive reasoning, and it is

thus an objective basis for analysing tectonic operations involving continental drift.

2.2.2: *CONTINENTAL RECONSTRUCTION IN THE SOUTH-WEST PACIFIC*

The starting point for the reconstruction which I have proposed (Griffiths, 1971a,b) is the fit of Australia and Antarctica, which has been independently computed at the 1000 fathom line (Sprall & Dietz, 1969) and at the 500 fathom line (Smith & Hallam, 1970). These fits are essentially identical, and appear to be supported by geological correlations between the Australian and Antarctic coasts (e.g: Plate XXIII in Bushnell & Craddock, 1970). Further geological investigations on the Antarctic coast are planned (Harrington, pers. comm; Oliver, pers. comm.) in an attempt to define more clearly geological correlations with Australia. In addition, the pattern of magnetic anomalies and transform faults south of Australia (Fig. 2.4, see also Sykes, 1970; McKenzie & Sclater, 1971; Weissel & Hayes, 1971, 1972) is consistent with reversal of drift to bring these continents into their suggested juxtaposition in the reconstruction, and hence this reconstruction appears to satisfy the requirements of Fig. 2.6.

I have previously discussed the delineation of the remaining fragments of continental crust, based on bathymetric and geophysical data (Griffiths, 1971a). I have generally taken the 2000 m line, smoothing out small irregularities (Fig. 2.7). The major submerged areas of continental crust are the Lord Howe Rise, Norfolk Ridge, Campbell Plateau and Chatham Rise (Fig. 2.7). The area immediately north of New Zealand has been reinterpreted

using more recent bathymetric data (Mammerickx, *et al.*, 1971), and a clearer definition of the Norfolk Ridge, Reinga Ridge and Three Kings Rise suggested, although there is only poor geophysical control for some of these features. The Dampier Ridge may also be a small continental sliver. Papua, the Louisiade Ridge and part of the Solomon Islands were also considered to be continental fragments (Griffiths, 1971a). The outline of these fragments used in the reconstruction is shown in Fig. 2.7.

Currently available geological and geophysical data (Figs. 2.3, 2.5) impose several constraints on the reconstruction. Stages leading to the final reconstruction now proposed (Fig. 2.8) can thus be summarised as follows:

- (i) The Australia-Antarctica fit previously discussed.
- (ii) Morphological fit of the Campbell Plateau against Antarctica, in a manner consistent with available marine magnetic profiler data (e.g: Pitman *et al.*, 1968; Christoffel & Ross, 1970; Falconer & Christoffel, in press).
- (iii) Closure of the central and northern Tasman Sea, bringing the Lord Howe Rise against eastern Australia with the Dampier Ridge as a small intervening fragment. This is consistent with the observed magnetic lineations in the central Tasman Sea (Hayes and Ringis, in press). Possible future work in this region might be directed towards gravity and magnetic studies on the Lord Howe Rise, to attempt to locate the extensions of the Sydney Basin and Great Serpentine Belt (Peel Thrust) of New

South Wales (see Figs. 2.5, 2.9) which would then serve as additional ties.

- (iv) Closure of the southern Tasman Sea. This entails reversing the 480 km offset on the Alpine Fault of New Zealand, and further "unbending" to straighten out reference lines in the New Zealand geosyncline (c.f: Carey, 1958; Fleming, 1970). This step leads to considerable simplification of the Mesozoic paleogeography of New Zealand (Griffiths, 1973; this thesis, Chapter 3). In addition much of the Tertiary structural development of New Zealand can be related to deformation during drift. Because of the severe Tertiary deformation which has occurred it is impossible to adequately delineate South Island. Thus the "gap" in the reconstruction (see Fig. 2.8) represents an area where most intense crustal shortening has subsequently occurred (see Chapter 3.3 for detailed discussion). The small fragments south and east of Tasmania are the South Tasman Rise and East Tasman Plateau, thought to be pieces of continental crust rifted from Tasmania (Houtz & Markl, 1972; Hayes & Conolly, 1972). Their incorporation in the reconstruction fills in otherwise small gaps.
- (v) Closure of the Bounty Trough, to bring the Chatham Rise against the Campbell Plateau, consistent with the suggested rift-like character of part of the Bounty Trough (Krause, 1966). Bollons Seamount fits in the

small intervening gap.

(vi) Closure of the basins north of New Zealand

to bring the Norfolk Ridge against the Lord

Howe Rise, the Reinga Ridge against the West Norfolk

Ridge, and the Three Kings Rise against the Norfolk

Ridge. The eastern part of North Island, New Zealand

then comes back into position as shown (Fig. 2.8). By

this stage the New Zealand Geosyncline has become an

essentially straight feature, bordering the Pacific

Ocean, continuous from New Caledonia through New

Zealand to the Chatham Islands and Antarctica.

(vii) Closure of the Coral Sea, and fitting the Louisiade

Ridge and part of the Solomon Islands in the remaining

gap.

Each stage of this reconstruction is now consistent, as far as data currently available, with the requirements of Fig. 2.6. The following general comments can be made:-

- (a) the reconstruction implicitly assumes that no movement has occurred between east and west Antarctica since at least the Paleozoic. I know of no evidence which invalidates this assumption.
- (b) there may be errors in the bathymetry, resulting in incorrect definition of the continental fragments. These errors may be significant in determining the *width* of long narrow ridges, but are not likely affect their *length*. In the reconstruction, the true lengths of the submarine ridges are preserved, and they fit

exactly as shown, provided that (a) is valid.

- (c) I will show later (Section 2.5) that, taking this reconstruction as a starting point, a model can be constructed which brings all the fragments to their present day position and in so doing simulates the magnetic anomaly pattern observed on the sea floor.

I will therefore adopt the reconstruction which I have proposed (Fig. 2.8) as a basis for further discussion of the south-west Pacific region in this thesis. Any additions or alterations to the input parameters of Fig. 2.6 which are required in the light of new data will necessitate re-evaluation of the reconstruction. This can be done objectively by following through the scheme of Fig. 2.6 below the point of input of the new data.

2.3: TECTONIC HISTORY PRECEDING UPPER MESOZOIC AND CENOZOIC CONTINENTAL DRIFT

The reconstruction shows the relative positions of the continental masses in the south-west Pacific as they were before they began to separate in the Upper Mesozoic. Hence any earlier geological data can now be re-plotted on the reconstruction, as a basis for study of the pre-drift tectonic evolution. Four broad divisions can be made:-

- (a) the Precambrian shield areas of Australia and Antarctica;
- (b) the Paleozoic orogenic terrains of eastern Australia, Antarctica, and a small area of New Zealand;
- (c) the Mesozoic sediments bordering the Pacific Ocean in New Zealand, New Caledonia and New Guinea;

- (d) the Mesozoic platform sediments which cover (a) and (b) in places (see Fig. 2.5).

In this thesis I am concerned mainly with the post-Paleozoic tectonic evolution, but will briefly discuss the Paleozoic features in so far as they form the earlier tectonic framework within which the Mesozoic and Cenozoic events have occurred. The history of the Paleozoic orogenic terrains involves interaction with the Precambrian basement, and hence (a) and (b) can be discussed together (Section 2.3.1). I will then consider briefly the evolution of the Mesozoic continental margin (c) in Section 2.3.2 and the platform cover (d) in Section 2.3.3.

2.3.1: PALEOZOIC TECTONIC EVOLUTION

Paleozoic rocks in the south-west Pacific region are mainly exposed in eastern Australia, where they comprise part of a large mobile belt, referred to here as the Tasman Orogenic Zone (after Solomon & Griffiths, 1972), but which has previously been called the Tasman Geosyncline. It is generally recognised that the Tasman Orogenic Zone is a complex orogenic belt, made up of several tectonic units which are distinct in both space and time. Much basic information is incorporated in the Tectonic Map of Australia and New Guinea (Geological Society of Australia, 1971).

Contemporaneous Paleozoic rocks are known in New Zealand (e.g: Grindley, *et al.*, 1959), and in Victoria Land, Antarctica (e.g: Gair, *et al.*, 1970). The trend of the Paleozoic rocks of eastern Australia strikes across the present coastline (Fig. 2.5) and hence it is highly probable that parts of the now submerged continental masses east of the Tasman Sea are also composed of

Paleozoic rocks. Although there is virtually no geological data available from these areas, their existence must still be taken into account in any interpretation of eastern Australian Paleozoic geology.

Recently attempts have been made to interpret parts of the Tasman Orogenic Zone in terms of plate tectonics (e.g: Oversby, 1971; Solomon & Griffiths, 1972; Schiebner, 1972; Heidecker, 1972). Excepting Solomon & Griffiths (1972) all of these attempts are basically similar in two respects, in that they propose gradual accretion of sediments and island-arc complexes onto the Australian Precambrian shield, and are confined to an analysis of the Australian mainland.

Solomon & Griffiths (1972) agree with the basic concept of continental accretion, but consider the whole south-west Pacific region of Gondwanaland. We suggest (*op. cit.*) that at least one large Precambrian "island" lay out in the Pacific Ocean well to the east, and during the evolution of the orogen these blocks have been carried westwards as ocean basins were consumed, and have subsequently collided with the main continental mass of Australia-Antarctica. Thus we interpret the Tasman Orogenic Zone as a complex of ortho- and para-tectonic belts (c.f: Dewey, 1969b; Dewey & Bird, 1970a). I tentatively consider that the Paleozoic tectonic framework of the south-west Pacific can be interpreted as in Fig. 2.9. At this stage it would be premature to attempt paleogeographic reconstructions for different periods. Interpretations are hindered by paucity of data, in particular poor stratigraphic control and lack of chemical data for the many volcanic rocks.

Direct evidence for the postulated Precambrian blocks is sparse, as the areas concerned are now almost totally submerged (Fig. 2.9). In Tasmania, Precambrian rocks occur east of the Dundas Trough (Figs. 2.5, 2.9), which is interpreted as the site of a consumed early Cambrian ocean basin (Solomon & Griffiths, 1972). Aronson (1968) concluded on the basis of Precambrian radiometric dates from detrital zircons, that Precambrian continental crust existed close to western New Zealand, which at the time was a considerable distance east of the Precambrian Australian-Antarctic shield (Fig. 2.9). Aronson & Tilton (1971) similarly suggest that Precambrian zircons found in New Caledonia were derived from possible Precambrian crust on the Lord Howe Rise to the west.

It is also important to consider the width of the Tasman Orogenic Zone in discussing its tectonic evolution. If the assumption is implicitly adopted (c.f: Oversby, 1971; Scheibner, 1972) that the whole width of the Zone, between the Australian/Antarctic shield and the Permian continental margin in New Caledonia and New Zealand, has formed during the Cambrian-Permian interval (about 350 MY) by eastward accumulation of sediments derived from the Australian-Antarctic shield, then there would seem to be considerable difficulties in deriving the vast amount of material required to produce the large areas of continental crust involved, particularly in the south where the zone is up to 1500 km wide.

Resolution of many of the problems of the Paleozoic tectonic evolution is hindered by general lack of adequate geological data,

and by the fact that large areas are now covered by water, ice, or platform cover. None of the syntheses so far proposed is likely to be substantiated as new data becomes available. Their main contribution at this stage is probably to define critical areas from which more information is required, and it is in this way that the plate tectonic hypothesis is making a contribution to the interpretation of the Paleozoic geology of the south-west Pacific.

It is possible to be more precise about the *termination* of the Paleozoic tectonic activity. The final stage in orogenesis of many areas of the world is the emplacement of "post-orogenic", high-level granite plutons, which are discordant to earlier structures. Radiometric age determinations on such plutons will then give an estimate of the end of orogenesis - i.e. the time of "cratonisation" of the area concerned. Granitic plutons occupy about 35% of the exposed Paleozoic rocks of eastern Australia (Solomon *et al.*, 1972) and many have been dated radiometrically (Fig. 2.10, based on Solomon *et al.*, 1972, Fig. 3). From this figure it can be seen that granite intrusions were largely emplaced during the Cambrian to Permo-Triassic, with occasional younger granites in the north. This distribution of ages suggests that cratonisation of the Tasman Orogenic Zone was largely complete by the end of the Paleozoic or earliest Mesozoic, and from thence on there is no evidence of major internal deformation within this craton until the Upper Mesozoic and Cenozoic episode of continental drift.

2.3.2: *THE PERMIAN-MESOZOIC CONTINENTAL MARGIN OF GONDWANALAND*

Orogenic activity persisted longest in the eastern part of the Tasman Orogenic Zone (see Fig. 2.11). The final deformation occurred during the Permian Hunter-Bowen orogenic movements, which were possibly related to the westward movement and collision of an older block now forming part of the Lord Howe Rise (Fig. 2.9). Much volcanism was associated with this phase during the Carboniferous-Permian in eastern Australia. The Pacific margin in New Zealand appears to have been a subduction zone during the Carboniferous (?) - Early Permian (see Chapter 3), and this zone may therefore have continued northwards along the margin (Fig. 2.9). Orogenic movements in eastern Australia were confined to a narrow zone by the Triassic (Fig. 2.11), in the region nearest to the continental margin (c.f: Solomon & Griffiths, 1972). Granite emplacement in the Permo-Triassic marked the final stages of the Hunter-Bowen orogenesis.

During these final stages, foredeep basins developed on the continental side of the orogenic belt (the Sydney and Bowen Basins). Subduction had ceased in New Zealand by the Middle Permian (Chapter 3) and the continental margin reverted to an "Atlantic-type" one. From thence until the mid-Jurassic the whole region was stable and little tectonic activity occurred.

During this time the Pacific margin was the site of deposition of a thick sedimentary pile, largely consisting of clastics derived from the continental foreland to the west. These sediments now outcrop in New Zealand, New Caledonia, New Guinea and on a few other small islands. The major outcrop of these continental margin sediments is in New Zealand (Chapter 3).

The succession in New Caledonia is remarkably similar,

and many units can be correlated with New Zealand (c.f: Lillie & Brothers, 1970, pp. 172-5). Similarities also exist with New Guinea (c.f: Lillie & Brothers, 1970, p. 177). The Chatham Islands are a continuation of the marginal sequence to the south (c.f: Hay *et al.*, 1970; Fleming, 1970, fig. 27), and possible equivalents also occur in Marie Byrd Land, Antarctica (c.f: Wade, 1970). The Pacific margin was thus a continuous but tectonically quiet belt of sediment accumulation, stretching at least 6000 km, during much of the Mesozoic.

Tectonic activity was renewed during the Mid-Jurassic, when a new episode of subduction commenced. An orthotectonic orogen developed as a consequence. This is well exposed in New Zealand (see Chapter 3) and in New Caledonia (Lillie & Brothers, 1970). The Cretaceous granites in eastern Queensland (Fig. 2.10) are perhaps related to this event, but otherwise its effects are not seen in eastern Australia.

2.3.3: PERMIAN AND MESOZOIC PLATFORM COVER

During the Permian the Sydney and Bowen Basins developed as flanking troughs on the continental side of the area of orogenic activity (Fig. 2.11). Sediments in these basins are generally shallow water marine or brackish (e.g: Sprigg, 1967), and were probably derived largely from erosion of the mountains to the east. Elsewhere on the craton, sediments are generally non-marine during the Permian, and glacial deposits extended over much of Australia (Fig. 2.11).

Very little deposition occurred in the Triassic, but by the

Jurassic and Cretaceous sediments are again more widespread in Australia, in-filling large shallow basins. In Antarctica the Beacon Group forms the platform cover, and is quite similar to the Australian succession and the Gondwanaland sequence in general. None of these platform sediments have been subsequently disturbed by significant orogenic movements.

By the Jurassic a rift valley was beginning to form between Australia and Antarctica (Fig. 2.12), and this received fairly rapid and thick clastic sediments derived from its flanks. Parts of this rift system later evolved into the present continental margins, whilst other areas became intra-cratonic, graben-type basins (see Chapter 4). Following cratonisation of the Tasman Orogenic Zone, the continental parts of Gondwanaland have remained as stable blocks, and all tectonic activity has been confined to the Pacific margin and later to the rifting of Gondwanaland.

2.4 SEA FLOOR SPREADING: PRECURSOR INTRACRATONIC EVENTS

The earliest signs of continental break up in the south-west Pacific are evident well before the onset of sea floor spreading. Along the southern Australian coast rifting began at least as early as the Jurassic (Griffiths, 1971b), suggesting that the precursor events to sea floor spreading may have occurred over a period of 100 MY or more. Prior to spreading the dominant tectonic regime was one of rift valley development, as very slow stretching of the craton occurred.

An evolutionary sequence in the development of a rift valley through to a continental margin has been recognised (e.g:

Carey, 1958; Dewey & Bird, 1970b; Griffiths, 1971b). It seems evident that the initial configuration of the rift valleys controls the later position of ocean basins, if the sequence evolves far enough, and it might be expected that pre-existing basement structures would exert some influence on the rift valley pattern. However, in my studies of south-east Australia (Chapter 4) I have not been able to establish any obvious relationship between the rift-forming faults and older structures on a regional scale, which suggests that although the initial rift pattern probably develops in response to stresses within the craton, basement structures have little influence, except perhaps locally, on its configuration.

The probable pattern of major rift valleys is shown in Fig. 2.12. Comparison of this figure with Fig. 2.9 clearly shows the lack of any obvious relationship between the rift system and Paleozoic structures. Extensive seismic reflection data from oil company work defines the northern side of the Australia/Antarctica rift, and the structures in the Bass Strait (c.f: Richards & Hopkins, 1969). Elsewhere a rift fault pattern has not been so clearly established. Houtz & Markl (1972) and Houtz (pers. comm.) indicate generally that the Antarctic margin south of Australia has rift-like characteristics, whereas the Tasman Sea, Ross Sea and Campbell Plateau margins are relatively steep and normal faults have not been observed. This may be a function of the amount of seismic work carried out, and the penetration achieved, rather than reflect absence of faults, as in general these early basement faults are covered by thick blankets of later sediments.

This rather irregular pattern of rift valleys (Fig. 2.12) is comparable to that of the African rift system of the present day. In the south-west Pacific some of the rifts have opened to eventually become ocean basins, whereas others have only opened a little. Most of the basin-forming structures of the southern Australian margin are faults which are *parallel* to the present continental margin, and which were only active *before* sea floor spreading began (see Chapter 4). Other authors (e.g: Le Pichon & Hayes, 1971; Francheteau & Le Pichon, 1972) have previously suggested that the orientation of transform faults controls the development of marginal basins. Results from south-east Australia, based on excellent seismic records, are however in conflict with this conclusion, as the ridge-transform system essentially post-dates the margin faulting. Major transform faults apparently occur where sharp bends in the *older* rift valley were situated (compare Figs. 2.4, 2.12).

Contemporaneous igneous activity also occurred during the early rifting stage. Two phases are recognised:-

- (a) Jurassic tholeiitic dolerites, which occur as sills and feeder dykes in Tasmania (Spry & Banks, 1963) and Antarctica (e.g: Gair *et al.*, 1970). These however have no apparent spatial relationship to the pattern of plate fragmentation, but may reflect early thermal instability beneath the region.
- (b) Upper Jurassic/Mid-Cretaceous high-potash calc-alkaline and shoshonitic rocks near to the Tasmanian coasts (Sutherland, in press, 1972) and in the Otway Basin

(Hocking, *in* Griffiths, 1971b). These are closely related spatially to the rift system, and may thus be tectonic equivalents to the volcanics in the African rift valleys (see Chapter 4).

Thus the early rift pattern is the dominant factor controlling transform fault development. I would emphasise again the timing involved, especially the slow initial development of rift valleys, which can take up to 100 MY or more. A more detailed analysis of these events is presented in Chapter 4.

2.5 SEA-FLOOR SPREADING:

EVOLUTIONARY MODEL FOR THE SOUTH-WEST PACIFIC

The basic tenet of the sea-floor spreading hypothesis is that magnetic anomalies are generated at centres of spreading (mid-ocean ridges), and are bilaterally symmetrical about these centres (Vine & Matthews, 1963). The anomaly patterns observed in different oceans are remarkably similar, and hence a "standard" time scale has been erected (Heirtzler *et al.*, 1968). Wilson (1967) introduced the concept of transform faults, which offset mid-ocean ridges and their associated magnetic anomaly patterns. Transform faults are parallel to small circles about the pole of rotation between two plates, and hence mapping of fracture zones (dead transform fault traces) provides additional control on the direction of relative motion between spreading continents.

Therefore, in theory, if a complete map of the ocean floor magnetic anomalies and fracture zones is available, then the

relative positions of continents at different times in the past can be deduced (c.f: Le Pichon, 1968). In practice, however, there are two complicating factors:-

- (i) the incompleteness of magnetic profiler data at the present time;
- (ii) the anomaly pattern only usefully extends back about 85 MY.

Thus in order to reconstruct former continental positions, a different approach must be adopted if there are insufficient data available from magnetic profiling alone, as is the case in the south-west Pacific.

2.5.1: BASIC PRINCIPLES OF MODEL CONSTRUCTION

The basic scheme for the construction of a sea-floor spreading model is illustrated as a flow chart in Fig. 2.13. The initial input has two components:- (i) the present positions of the continents; (ii) the available magnetic data between the continents. If (ii) covers the whole intervening area of ocean floor, then a rigorous geometrical synthesis can immediately be made.

In the more usual case of this requirement not being satisfied, then a different scheme must be adopted. Firstly a *partial synthesis* can be made from the available data. Then an additional constraint is used - the morphological pre-drift fit of the continents, derived as outlined in Section 2.2 (see Fig. 2.6). Thus there are now *two* end-points, with intervening constraints based on magnetics, from which to construct a sea floor spreading model, rather than only being able to work backwards

from the present day situation. The proposed model is checked for consistency with the marine magnetics, and thus a synthesis of the sea floor spreading consistent with the available data is made.

This synthesis is in turn *predictive*, and so a theoretical magnetic anomaly map can be drawn, and checked against data acquired later. If eventually magnetic mapping coverage is completed, then a rigorous geometrical synthesis can be made, but until this is possible, the scheme outlined allows progressive evaluation and modification of the synthesis compatible with new data. [Note also that the predictive map can be tested against sea floor ages obtained from deep-sea drilling]. Ultimately the marine magnetic data alone should be sufficient to derive a pre-drift reconstruction, which, if the basic assumption of the validity of the sea floor spreading hypothesis is correct, should be similar to the morphological fit proposed.

2.5.2: CONSTRUCTION OF THE MODEL FOR THE SOUTH-WEST PACIFIC

The extent of available magnetic data in the south-west Pacific has been reviewed in Section 2.1, and summarised in Fig. 2.4. The magnetic data define the spreading history between Australia and Antarctica (c.f: Weissel & Hayes, 1971, 1972). Only the northern side of the Pacific-Antarctic Ridge has been adequately mapped, and anomalies have also been identified in the central Tasman Sea. Using the principles discussed above (Section 2.5.1), the basic approach to construction of a sea floor spreading model for the south-west Pacific region is therefore as outlined in Fig. 2.14. The analysis of Weissel & Hayes (1972) is adopted as part of

the model. The remaining magnetic data provide constraints on the rest of the model, and the morphological reconstruction of Fig. 2.8 is also used.

Construction of the actual model is a multi-stage procedure, which I have outlined using a flow chart (Fig. 2.15). This has been broken into three stages for discussion. The first two parts (Fig. 2.15 a, b) reflect the major reorientation of plate geometry which occurred about 55 MY ago. These two stages are then integrated (Fig. 2.15 c), and the results of the model study are presented as a series of paleotectonic maps at 10 MY intervals.

The mechanical operations were carried out using a 1:15 million Lambert's equal-area azimuthal projection, centred at 60°S, 160°E, as a base. This has only slight angular distortion away from its centre (less than 5°), and gives a good visual picture of the region. However, operations involving rotations were checked on a stereographic projection, and the results replotted on the equal-area projection as appropriate.

The basis of the model is the progressive relative shifting of the continental blocks, at each stage checking for compatibility with magnetic data in the intervening oceans. Thus decisions (diamonds on the flow charts) have to be made, and the model continuously adjusted so that it is consistent with the constraining data at each stage. In practice, the diagrams were constructed using tracing paper overlays, and adjusting at each stage until agreement with magnetic data was achieved. Availability of more data, and the use of computers to manipulate the blocks, may

eventually allow the model to be improved, but is unlikely to affect the major conclusions significantly.

Relative shifts between two blocks are conveniently recorded by specifying a *centre of rotation* (abbreviated as RC) for the motion. The centre of rotation is determined by holding one block fixed with respect to the present latitude-longitude co-ordinate system, and then obtaining the centre of rotation which gives the relative motion of the other block, over a specified time interval. Thus a centre of rotation is purely a numerical expression of the total motion specified, and is different from a *pole of rotation* (abbreviated as RP) which is the actual pole of instantaneous movement between two plates. I use the definitions of Harrison (1972), who discusses the differences in these two terms in more detail, and points out that the two are not synonymous unless it can be shown that the pole of rotation has remained fixed relative to both plates through time.

The first stage of the model (Fig. 2.15a) relates to the early opening of the South Pacific Ocean and Tasman Sea. Hayes & Ringis (in press) show that the central Tasman Sea evolved between 77 and 60 MY ago, and hence the relative positions of Australia, Antarctica and the Lord Howe Rise at 60 MY can be plotted (Fig. 2.16). Using the reconstruction of Fig. 2.8, the centre of rotation for the Lord Howe Rise relative to Australia over the interval 80 - 60 MY is determined to be 145°E , 15°S . Next it is assumed that the Lord Howe Rise and Campbell Plateau remained in the same relative positions from 80 - 60 MY. Using the centre of

rotation established, the movement of the whole Lord Howe Rise/Campbell Plateau block is derived, (Fig. 2.16). Small circles drawn about the rotation centre are in good agreement with the trends of the observed fracture zones west of the Campbell Plateau (see Fig. 2.16), and hence the assumption that the Lord Howe Rise and Campbell Plateau were a single rigid block during this stage appears justified. It is therefore possible to construct a predictive anomaly map for this interval (Fig. 2.17). Control on timing is based on the compilation of magnetic data in Fig. 2.4. When these data are extrapolated across the ridge, the "mirror image" is a good fit in the space available (Fig. 2.17).

The second stage of the model (Fig. 2.15b) relates to the spreading from 60 - 0 MY. The centre of rotation of the Campbell Plateau relative to Antarctica from 60 - 0 MY is determined as 120°E , 80°S . This point is close to the present instantaneous rotation pole of Christoffel (1971), but differs by some 10° in latitude from the present pole of Le Pichon (1968). For the moment I will use this centre, and discuss the possible causes of the differences later. Thus again small circles are drawn about this centre (Fig. 2.18), and compared to observe magnetic trends. In this case published data on magnetic lineations are insufficient to clearly define zones south of the Campbell Plateau, but earthquake epicentres, presumably lying on transform faults, are aligned roughly parallel to the small circles (see Fig. 2.18). Further east, Herron (1971) considers that the Eltanin Fracture Zone seems more consistent with the Le Pichon pole for 0 - 10 MY,

but that its earlier trend indicates a pole at 62°S , 165°E . This combination however leads to large discrepancies in fit when used to shift the Campbell Plateau back towards Antarctica from its position in Fig. 2.16. Chase (1971) breaks the Campbell Plateau-Antarctica motion into six stages each about 10 MY long, and derives a separate rotation centre for each stage. Use of these centres brings the Campbell Plateau into a similar orientation to that in Fig. 2.16, but allows for about 300 km less opening in the south Tasman Sea.

The cause of these discrepancies in different determinations of the rotation centres and poles is due to their close proximity to the information from which they are derived, such that small differences in fracture zone orientation lead to large errors in pole determinations. The problem is unlikely to be resolved until more data are available, and hence in this analysis I will use the total centre of rotation determined (80°S , 120°E), whilst bearing in mind the need for future modification of the model when the full details of the spreading pattern for 60 - 0 MY is ascertained.

Using this rotation centre, it is therefore possible to construct a predictive anomaly map for the interval 60 - 0 MY (Fig. 2.19). Time control is taken from Fig. 2.4. Any migration of the rotation pole which has occurred during this interval will have the effect of changing the apparent orientation of anomalies and fracture zones, particularly near to the centre. Thus the "Pacific-Antarctic Fracture Zone" (Christoffel & Falconer, 1972) could equally well be a series of *en echelon* short ridge segments and transform offsets.

The model which I have constructed is at this stage consistent with the available published data. Discrepancies may be inherent due to migration in time of instantaneous rotation poles, and to general lack of data. The two areas in particular from which more data are needed are towards the "apex" of the Pacific-Antarctic segment from 60 - 0 MY, and off the Antarctic coast from 80 - 60 MY. When this data becomes available, the model will need modifications, but these are not likely to be major.

The final stage (Fig. 2.15c) is to integrate the analyses of the 80 - 60 MY and 60 - 0 MY stages, and the established Australia-Antarctica spreading history (from Weissel & Hayes, 1972). This is done by plotting the positions of Australia-Lord Howe Rise relative to Antarctica (Fig. 2.20, from Fig. 2.17), and of the Campbell Plateau relative to Antarctica (Fig. 2.21, from Fig. 2.19). By combining these two plots, I have thus derived the basis of sea floor spreading model which is illustrated in 10 MY stages in Figs. 2.22 - 2.30. A predictive anomaly map for the interval 80 - 0 MY (Fig. 2.31) is constructed by combining these 10 MY stages and the data from Figs. 2.17 and 2.19. This sequence is described in the next section, following which I will discuss the Australian-Pacific plate boundary in more detail.

2.5.3: OPENING OF THE SOUTH PACIFIC OCEAN, TASMAN SEA AND SOUTH-EAST INDIAN OCEAN.

The sequence of 10 MY stages (Figs. 2.22 - 2.30) is largely self-explanatory, and the following discussion is intended to amplify some of the major points and provide continuity. In each

of the figures Antarctica is kept in a fixed orientation, and the relative positions of the other blocks and plate boundaries are shown. Segments of the ridge system, and the previous 10 MY increment of crustal growth, are indicated. Detailed discussion of New Zealand and of the area east of the Lord Howe Rise is presented later.

The earliest episode of sea-floor spreading in the south-west Pacific related to the disintegration of Gondwanaland was a northward opening between the Campbell Plateau and Antarctica, about a pole of rotation in the vicinity of 145°E , 15°S , at an angular rate of about 11×10^{-7} degrees/year (Figs. 2.22 - 2.24). This was probably a continuation of an active spreading ridge in the Pacific Ocean. East of the Eltanin Fracture Zone the history is rather complex (Herron, 1971), but on the ridge segment to the west, which continues up into the Tasman Sea, the anomalies suggest an "unzipping" or diachronous opening. Thus Falconer (1972) recognises the oldest anomaly south of the Campbell Plateau to be 83 MY, and Hayes & Ringis (in press) recognise anomalies back to 77 MY in the central Tasman Sea.

There is little evidence for rift-margins associated with this opening, but this may be due to the amount of seismic reflection work carried out, rather than the real lack of structures (see Chapter 4). During the early spreading phase it is probable that the East Tasman Plateau and Dampier Ridge were partly rifted from Tasmania and the Lord Howe Rise respectively, and left behind as small slivers of continental crust.

The youngest central Tasman Sea anomalies are about 60 MY (Hayes & Ringis, in press) and the profiles are symmetrical about a deeply submerged and sediment-covered ridge (Hayes & Ringis, in press; Symonds, pers. comm.). Farther north a line of seamounts may perhaps be related to spreading in the narrower part of the Tasman Sea.

Several fracture zones also relate to the opening phase, including those adjacent to the Campbell Plateau (Christoffel & Falconer, 1972), in the Tasman Sea (Hayes & Ringis, in press), and at the northern termination of the Lord Howe Rise (Symonds, pers. comm.). Minor rifts during this spreading phase may have also formed the Bounty Trough, Bellona Gap and the incipient Emerald Basin.

Spreading in the Tasman Sea would presumably have continued in a similar way, except that around 60 MY the south-east Indian Ridge between Australia and Antarctica became active. The Australia/Antarctica "crack" had been developing since the Jurassic (Section 2.4) without becoming a spreading ocean. The reasons for such sudden changes in spreading patterns must be sought in a world-wide study of plate movements, which is beyond the scope of this thesis. However, McKenzie & Sclater (1971, p. 507) indicate that a gross readjustment of the spreading pattern in the Indian Ocean occurred at about this time, with cessation of spreading on the Carlsberg Ridge, followed closely by initiation of spreading at the south-east Indian Ridge. This may have been related to the collision of India with Asia, and formation of the Himalayas (Le Pichon & Heirtzler, 1968, p.2115).

This ridge split through from the west (see Chapter 4), and followed the sharp bend of the earlier rift to the south of Tasmania by developing a series of transform faults and short ridge sections (Fig. 2.25). Thereafter it intersected the Pacific-Antarctic spreading axis at an angle of about 80° , resulting temporarily in a highly unstable plate geometry (compare Figs. 2.24, 2.25). Resolution of this situation occurred by cessation of spreading in the Tasman Sea, and breakthrough of the south-east Indian Ridge to the east. However, the Pacific spreading pattern could not totally adjust to this new direction, and exerted a restraint such that a new "compromise" plate geometry became established. Corresponding changes in the Pacific spreading pattern are also recorded at about 65 - 50 MY (e.g: Herron & Hayes, 1969; Christoffel & Ross, 1970; Christoffel & Falconer, 1972).

The period of readjustment would thus have been between about 60 - 55 MY, and hence a very confused anomaly pattern might be expected south of the Campbell Plateau and in the south Tasman Sea during this interval. The earlier two-plate (Australia/Antarctica, Pacific) geometry was thus replaced by a three-plate system (Australia, Antarctica, Pacific), in which three rotation poles are now needed to describe relative plate motions, rather than one as previously.

The motion of Australia from Antarctica follows a relatively simple pattern (Figs. 2.25 to 2.30). The oldest magnetic anomalies are probably 55 MY (Weissel & Hayes, 1971, 1972), which corresponds to early Eocene. The rifted nature of the southern

Australian continental margin is clearly shown on seismic records (see Chapter 4), which indicate that fault movement ceased during the Eocene (Griffiths, 1971b), and thus the transform faults can have had little if any controlling influence on the structural development of the marginal basins. Initially the offset of the ridge was perhaps accomplished by many small transform faults, but such a system is probably unstable (c.f: Vogt, *et al.*, 1969) and would soon resolve into the several large fracture zones now observed (Hayes & Conolly, 1972).

The northern limit of oceanic crust generated at the south-east Indian Ridge across the Tasman Sea is marked by a change in sea floor topography, sediment thickness and magnetic character (Hayes & Conolly, 1972; Houtz & Markl, 1972; Weissel & Hayes, 1972). Proposed JOIDES holes (see Fig. 2.5) should provide much more data on the age and nature of the Tasman Sea floor in this region.

I have previously shown (Fig. 2.18) that the Pacific/Antarctic plate motion (i.e: Campbell Plateau relative to Antarctica) from 60 - 0 MY is satisfactory summarised by rotation about a centre $120^{\circ}\text{E}, 80^{\circ}\text{S}$, although the actual rotation pole may have wandered over this period. The pole is relatively close to the Campbell Plateau, and hence apparent movement of the Plateau with respect to Antarctica is an anticlockwise rotation of about 40° (Fig. 2.21). By integrating Figs. 2.20 and 2.21, the probable history of the Australian/Pacific plate boundary from 60 - 0 MY can now be determined from the model.

2.6: THE AUSTRALIAN/PACIFIC PLATE BOUNDARY

The Australian-Pacific plate boundary runs northward from a triple point at the south-east Indian Ridge - Pacific/Antarctic Ridge junction, along the Macquarie Ridge, and the Alpine Fault in New Zealand, and splays into the Kermadec Trench (Figs. 2.1, 2.3). The Kermadec Trench is continuous with the Tonga Trench, and then the boundary links up with the Eurasian/Australian boundary in a very complex zone of interaction through Indonesia.

Griffiths & Varne (1972) made a preliminary attempt to construct a sea-floor spreading model for the south-west Pacific, including the Macquarie Ridge/Alpine Fault system, which we interpreted as being a major shear along which uplift had later occurred. New data is now available which necessitates some modifications being made to this model. It is now possible to discuss the evolution of the whole boundary in detail, based on the sea-floor spreading model which I have constructed in this thesis. The basic steps in this scheme are outlined in Fig. 2.32.

The actual boundary is conveniently considered in three parts. In this section I will concentrate mainly on the parts to the south and north of New Zealand, essentially where the boundary propagates through oceanic crust. Detailed discussion of its course through New Zealand will follow in Chapter 3.3, as part of an overall tectonic synthesis of New Zealand geology.

2.6.1: THE MACQUARIE RIDGE COMPLEX

The Macquarie Ridge has been variously interpreted as a mid-ocean ridge (Ewing & Heezen, 1965), an island arc (Summerhayes, 1969) or a shear zone (Houtz *et al.*, 1971). Other authors (e.g: Le Pichon, 1968; Christoffel, 1971; Hayes & Talwani, 1972; Hayes, *et al.*, 1972; Weissel & Hayes, 1972) have attempted to calculate the present vectors of relative plate motions in the south-west Pacific, and generally conclude that the Macquarie Ridge is currently a complex zone with discrete segments of extension, shear and compression. There are significant discrepancies between the vectors calculated by these authors, which are due to the close proximity but imprecise definition of the Australian/Pacific and Australian/Antarctic poles of rotation. Thus slight variations of the pole positions will significantly alter the vector orientations near to the pole (c.f: Fig. 2.3). The Tonga/Kermadec Trench is further from the area of the pole, and all models are similar in indicating subduction. Hayes & Talwani (1972) give a comprehensive account of the Macquarie Ridge and discuss these problems in detail.

All of these motion studies have generally only been extrapolated back about 10 MY, and thus have not considered the earlier evolution of the plate boundary. The model which I have constructed above for the past 80 MY allows such an analysis to be attempted, as outlined in Fig. 2.32. The relative positions of the Campbell Plateau and the Lord Howe Rise in 10 MY stages from 60 - 0 MY previously determined are shown in Figs. 2.24 to 2.30. The relative motions between the Campbell Plateau and the Lord Howe

Rise were determined from these figures by superimposing the outlines of the Lord Howe Rise and Australia on successive 10 MY pairs, and then obtaining the Campbell Plateau/Lord Howe Rise rotation centre for each stage. These centres were then plotted with respect to Australian coordinates (Fig. 2.33). A curve drawn through these points probably approximates to the migration path of the *pole of rotation* between these two blocks.

Intermediate positions of this pole were then plotted on each 10 MY map, and small circles drawn about these poles in order to derive the approximate orientation of the movement vectors at the boundary (see Figs. 2.24 to 2.30). The pole at 0 MY is thus at about 160°W, 58°S.

The 10 MY increments of offset between the Campbell Plateau and Lord Howe Rise are also measured (see Fig. 2.33), thus giving an indication of the timing of movements on the Alpine Fault. The total offset amounts to about 900 km along the Macquarie Ridge. This is about 400 km more than the observed displacement on the Alpine Fault (Wellman, *in* Benson, 1952), the excess probably being accounted for by internal deformation in New Zealand (see Chapter 3.3).

The Australian/Pacific plate boundary south of New Zealand appears to have gone through progressive stages of development. After the initiation of the new sea floor spreading pattern at 60 MY the spreading rate on the Pacific/Antarctic Ridge was considerably less than on the south-east Indian Ridge, although the spreading axes are roughly parallel (compare anomaly spacings and orientation in Fig. 2.4). Thus the Campbell Plateau is rotating anticlockwise

relative to the Lord Howe Rise. This system implies that at least one transform fault on the ridge system must be progressively lengthening, which is an unstable situation over any period of time (see Fig. 2.34a). Restoration of stability can only be achieved if one or both ends of the transform propagate beyond the ridge ends, the simplest solution being that shown in Fig. 2.34b, where one ridge-transform junction becomes a ridge-fault-transform triple junction.

The solution in the south-west Pacific is however more complex. A triple junction developed at the northern end of a transform fault on the Pacific/Antarctic ridge, but the required offset was taken up at first by propagation of oblique extension across a north-easterly trending zone. This is shown diagrammatically in Fig. 2.34c (compare with Figs. 2.25, 2.26). The "oblique spreading zone" is identified as the Emerald Basin. Hayes & Talwani (1972) show NNE trending magnetic anomalies in the Emerald Basin just east of the present Macquarie Ridge, which are consistent with the above model. Since these anomalies are undated, it is not possible to determine the location of the spreading centre in the Emerald Basin. I have only diagrammatically suggested that it was along the northern edge of the basin, but this is probably the most likely location as the northern margin appears to have progressively evolved into a shear zone (see below).

Spreading in the Emerald Basin decreases towards the north-^{east}~~west~~, as the trend of the incipient plate boundary swings round to a northerly direction and becomes a zone of dextral

transcurrent offset (see Fig. 2.34c). These motions correspond to those indicated by small circles about the derived rotation poles (see Figs. 2.25, 2.26). The triple point appears to migrate west along a section of ridge which diminishes in length, whilst the south-east transform fault is growing and its north-easterly extension has both extensional and transcurrent components of motion. This involves a rather complex, but nevertheless stable, triple point (Fig. 2.34c).

By 30 MY the Campbell Plateau/Lord Howe Rise rotation pole appears to be migrating rapidly south (Fig. 2.33). Small circles to this pole (compare Figs. 2.26 to 2.28) thus now become increasingly more parallel to the northern margin of the Emerald Basin, and so transcurrent rather than extensional motion becomes dominant. Available data are insufficient to indicate whether this was a gradual change or a sudden jump, but the result was that the northern margin of the Emerald Basin (roughly parallel to the line of the future Macquarie Ridge) became a dextral transcurrent fault (Fig. 2.28). The triple point at this stage therefore becomes a ridge-fault-fault one, and following the stability criteria of McKenzie & Morgan (1969), the ridge axis must now equally bisect the angle between the two faults, and the ridge segment itself will continue to diminish in length (see Fig. 2.34d).

Increasing relative rotation of the Campbell Plateau led to significant dextral offset against the Lord Howe Rise, and hence the Alpine Fault through New Zealand must have become a discrete fault about this time. (The geological consequences of

this model are discussed in Chapter 3.3). With rotation continuing to the present (Figs. 2.29, 2.30), the total displacement on the Macquarie Ridge plate-boundary has been about 900 km. The rotation pole has continued to migrate south slowly (Fig. 2.33), with corresponding changes in vector orientation, so that at present the southern end of the Macquarie Ridge may be an incipient small oblique subduction zone (see Fig. 2.30), with a triple point developing as shown in Fig. 2.34e.

Subduction is apparently beginning in Fiordland, South Island (c.f: Christoffel & Van der Linden, 1972) where a reasonably well-defined Benioff zone dips steeply south-east to about 150 km (Smith, 1971). Uplift of Macquarie Island has also been a recent and continuing event (Varne *et al.*, 1969). Williamson (1972) considers Macquarie Island (and hence by inference part of the Macquarie Ridge) to be a buckled-up part of the Australian plate, resulting from compression across the plate boundary. He identifies magnetic anomaly 8 (29 MY, Upper Oligocene) running across the northern end of the island.

The latter stages suggested by the model are quite compatible with previous analyses which cover the last 10 MY or so (c.f: Christoffel, 1971; Weissel & Hayes, 1972; Hayes & Talwani, 1972). The topographic expression of the Macquarie Ridge is probably due to a slight pole shift, such that the vectors along the plate boundary are no longer quite parallel to it, and some compression is occurring (c.f: Williamson, 1972). Speculating on the future, it seems that either further compression and more subduction may occur (but still with a dominant shear

component), or perhaps more likely that the juxtaposition of continental masses across the plate boundary in New Zealand will inhibit formation of a major subduction zone and constrain any great movement of the rotation pole, thus keeping the movement vectors fairly parallel to the plate boundary.

On the basis of this model I have constructed a predictive plot of the magnetic anomalies which should be observed (Fig. 2.31). Because there are still uncertainties in the determination of the present rotation poles and their past movement paths, this plot is unlikely to prove accurate in detail. However, it serves to indicate the degree of complexity which might be expected in the magnetic anomaly pattern. Detailed resolution of the magnetics will need very closely spaced tracks, particularly near the Macquarie Ridge triple point and around the Pacific/Antarctic Fracture Zone. As such new data becomes available, it can easily be fed into the flow schemes which I have developed for the model construction (Fig. 2.15), at the appropriate point, and then all subsequent stages in the scheme can be reworked to include the new data and upgrade the model.

2.6.2: *THE PLATE BOUNDARY NORTH OF NEW ZEALAND.*

In the previous section I have implicitly assumed that the Macquarie Ridge complex is continuous with the Alpine Fault in New Zealand, and hence the dextral offset on the Macquarie Ridge must be transmitted along this fault. This is consistent with observations on the Alpine Fault (c.f: Wellman, *in* Benson, 1952; Wellman, 1956), and allows a chronology of movements on this fault

to be determined from the sea-floor spreading model. The Alpine Fault is clearly defined in South Island, where it dextrally displaces Paleozoic and Mesozoic sequences by 480 km (Wellman *in* Benson, 1952; Wellman, 1956). It was shown above (Section 2.2.2) that to derive the Gondwanaland reconstruction, both this 480 km, and a further amount due to related Tertiary bending, must be reversed. These movements are discussed later (Chapter 3.3, see especially Table 3.1). However, the implication is that the plate boundary must continue northwards through New Zealand.

Suggate (1963, fig. 5) projects the Alpine Fault through North Island, east of the Taupo Volcanic Zone. [In fact the movement may be dissipated through several faults - see Chapter 3.3 - but with the same net result]. Various analyses of the Australian/Pacific plate boundary have adopted this postulated extension of the Alpine Fault, linking it with the southern end of the Kermadec Trench to form a continuous plate boundary (e.g: Summerhayes, 1967, 1969; Isacks, Oliver and Sykes, 1968; Le Pichon, 1968; McKenzie and Morgan, 1969; Karig, 1970 a,b, 1971; Christoffel, 1971). However, these analyses deal with only the most recent part of the boundary's evolution. As previously demonstrated (Section 2.6.1), the *pole of rotation* for the Australian/Pacific boundary appears to have migrated rapidly during the last 50 MY, and hence the position of the plate boundary may also have changed considerably during this time.

A possible pre-drift extension of the future Alpine Fault line is suggested by the reconstruction (Fig. 2.8). When the Campbell Plateau and Lord Howe Rise were replaced alongside Australia/Antarctica, the

eastern part of North Island had to be displaced relative to the western half by about 900 km. Closure of the New Caledonia and Norfolk Basins was achieved by offset of the Norfolk Ridge at about 30°S, and shift of the Three Kings Rise into the Norfolk Basin. The simplest solution to obtain these required shifts would be transcurrent faults along the northern and southern margins of the Norfolk Basin, with the southern one continuing south into New Zealand and west to cut the Norfolk Ridge (See Fig. 2.8).

Inspection of the bathymetry of the area (Van der Linden, 1967; Mammerickx *et al.*, 1971) reveals two fracture zones (Vening-Meinesz and Cook Fracture Zones) which may correspond to the transcurrent faults suggested above. Thus it is postulated that the Alpine Fault originally continued through New Zealand, to link up with the Vening-Meinesz Fracture Zone. Evidence for this connection should occur in the area of the Bay of Plenty, but in this region the Quaternary Taupo Volcanic Zone has probably disrupted any earlier structures. However, using this suggestion, it is shown below how a model for the plate boundary north of New Zealand, as far as New Guinea, can be constructed which appears consistent with available geological and geophysical data.

Chase (1971) has also attempted an analysis of part of this boundary, around the Fiji Plateau area, at the northern end of the Tonga Trench, using computer models to shift plate boundaries in time in a way which appears to simulate observed bathymetric and geological features. Whilst his interpretation of present day plate boundaries may well be correct, he assumes (p.3105, *op.cit.*) that both the Tasman Sea and South Fiji Basin opened prior to 45 MY. I would question the second of these assumptions, and ~~the~~ hence ^{the} validity of the analysis. His interpretations seem to be inconsistent with the overall history of the plate boundary north of New Zealand, and

difficult to reconcile with other geological data from the Solomon Islands and New Caledonia, which are outside the area discussed by Chase (1971). In areas with good magnetic control (e.g: south of Australia), computer studies are perhaps the best approach to reconstructing former positions, but in areas such as Melansea, geological constraints are far more important. It would probably be difficult at this stage to incorporate all such constraints into any purely computer-generated models, and in such areas reconstruction by "eye" is likely to be the best method.

Chase (1971) works only backwards from the present day in establishing former positions. In the following analysis however, I have used both the present positions *and* the positions of the continental fragments as they were at 80 MY. Thus the problem now is to determine an evolutionary sequence from 80 - 0 MY which simulates the observed features of the region. In the absence of adequate magnetic data for chronological control of movements, geological information is of prime importance. Although the resulting model is therefore less quantitative, it still appears consistent with observations, and is also predictive. The model is illustrated in 10 MY stages, which have been incorporated into Figs. 2.22 - 2.30.

The centre of rotation to open the Coral Sea is near to its apex, at about 145°E , 12°S . This centre is very close to that previously and independently derived for the Tasman Sea from 80 - 60 MY (145°E , 15°S). The proximity of these two centres is probably more than coincidence, and implies that the opening of the Coral and Tasman Seas were related and simultaneous events

(Figs. 2.22 - 2.24). The angular opening of the Coral Sea is about 25 degrees, compared to 23 degrees for the Tasman Sea. JOIDES (1972) holes 209 and 210 indicate a minimum Eocene (50 MY) age for the Coral Sea. The Coral Sea sphenochasm is offset from the Tasman Sea by a fracture zone which forms the northern limit of the Lord Howe Rise (Figs. 2.22 - 2.24). In the model the Solomon Islands are also moved about this rotation centre, to the position indicated in Fig. 2.25.

Thus the early (80 - 60 MY) history of the south-west Pacific region can be satisfactorily synthesised by sea-floor spreading about a single rotation centre describing the motion of the continental fragments from New Guinea to New Zealand. During this stage the Pacific/Antarctic plate boundary is thus the spreading axis in the Tasman Sea, and all the continental fragments to the east lie on the Pacific plate.

The geometry of sea floor spreading changed rapidly about 60 - 55 MY ago, as already outlined. Differential separation rates of the Australian and Pacific plates from Antarctica resulted in the establishment of an Australian/Pacific pole near to New Zealand. At 50 MY, small circles about this pole indicate convergence between the Australian and Pacific plates (Fig. 2.25). In the south this was accomodated largely by transcurrent movement along the Macquarie Ridge/Alpine Fault. North of New Zealand small circles are oblique to the old plate boundary in the Tasman Sea, and a new tectonic regime was established. The plate boundary appears to have "jumped" eastwards to a position along the outer edge of the

continental fragments, where the plate convergence was accommodated by development of a subduction zone. This position was presumably more favourable for subduction than within the Tasman Sea, where in any case only a limited amount of subduction could have occurred before the Lord Howe Rise would have collided with the Australian coast.

Initially this subduction zone was close to the continental edge (Fig. 2.25). Vectors of motion are small near to New Zealand, and increase northwards along the boundary further away from the pole. Assuming that the same rotation pole can be used for the whole boundary, the vectors were at first roughly normal to the continental margin as far as New Guinea. (Fig. 2.25).

After the Upper Mesozoic episode of subduction at the Gondwana margin (Section 2.3.2), it was a passive continental edge until subduction was renewed in the early Tertiary. Evidence for this event is seen in the geology of several exposed parts of the outer margin such as in New Caledonia, the Solomon Islands and New Guinea. The initial subduction zone dipped west, under the continent. In its early stages of development, ophiolite sequences, representing oceanic crust, were emplacement by "obduction" (Coleman, 1971), or overthrusting from the east, with associated volcanic activity (Fig. 2.25).

In New Caledonia the extrusion of Late Eocene and Oligocene basalts was followed by emplacement of peridotites and serpentinites along major thrusts (c.f: Lillie & Brothers, 1970), accompanied by blueschist metamorphism in the underlying rocks (c.f: Coleman, 1971). Rapid uplift occurred after the Eocene

(Lillie & Brothers, 1970). In the Solomon Islands, Upper Eocene-Oligocene lavas (andesites and pillow lavas) intrude and partly cover the older basal complex of the central province (P. Coleman, 1966, fig. 2) (This spatial relationship is suggestive of a *south-west* dipping Benioff zone in the Oligocene). In Papua, a large ophiolite complex was emplaced from the north, probably in the Eocene, accompanied again by blueschist metamorphism (Davies, 1968; Davies & Smith, 1971).

Once established, this subduction zone appears to have maintained fairly rapidly eastwards (Figs. 2.26 to 2.30) creating marginal basins behind the trench-arc as has been suggested by Karig (1971). Any offset in this trench would most probably occur as a transform fault parallel to a small circle about the rotation pole, and such an offset may thus have defined the later location of the Hunter Fracture Zone (see Figs. 2.26, 2.27). In the model the New Hebrides arc is interpreted as being initially an eastward migrating island arc above a *west* dipping Benioff zone, which migrated eastwards more slowly than the Tonga arc, with offset occurring on the Hunter Fracture Zone (Figs. 2.26, 2.27).

The New Hebrides at present are located above an east-dipping Benioff Zone, and earthquakes are recorded to a depth of about 600 km (Isacks *et al.*, 1968). Thus at least 1000 km of oceanic crust have been consumed, and the arc must originally have been well to the east of its present position. The New Hebrides arc could therefore either have developed with its present polarity at some distance from the continental edge, or have migrated out from the edge and then, as a result of a

change in polarity of subduction from west to east-dipping, have been carried back west towards New Caledonia as suggested here (Fig. 2.26 - 2.30). There is some evidence for such an early trench located east of the New Hebrides, as ophiolites occur in the eastern belt and associated calc-alkaline volcanics to the west (Mitchell & Warden, 1971, fig. 5). These authors have likewise speculated on the possibility of a change in polarity of the New Hebrides and Solomon arcs.

Migration of the early trench and arc eastwards would thus lead to formation of the South Fiji Basin during the period 50 - 30 MY (Figs. 2.25 - 2.27). JOIDES (1972) hole 205 indicates a minimum Upper Oligocene age for the sea floor in the eastern South Fiji Basin, which is consistent with this model. During this period a secondary spreading system also appears to have developed within the Australian plate. The motions of the Norfolk Ridge and Three Kings Rise, and hence the opening of the New Caledonia and Norfolk Basins, can be simulated using a rotation centre at about 167°E , 38°S (with respect to Australian co-ordinates). The Cook and Vening-Meinesz Fracture Zones are parallel to small circles about this centre, as is the Alpine Fault line (see Figs. 2.26 to 2.29). Thus in the model, the Alpine Fault and Vening-Meinesz Fracture Zone can be treated as a continuous feature, supporting the previously postulated connection between these features (e.g. Fig. 2.8). This spreading had probably begun before 40 MY. JOIDES (1972) hole 206 records Eocene deep-sea sediments in the New Caledonia Basin. The geometry of this system is complex and has not been quantitatively analysed. It is shown diagrammatically in Fig. 2.35, which illustrates how

spreading in the New Caledonia Basin could have been related to the dextral displacement on the Alpine Fault and eastward migration of the Tonga Trench. This opening probably proceeded in several stages. First the New Caledonia Basin opened (Fig. 2.26). The Norfolk Ridge is offset on the Vening-Meinesz Fracture Zone, so that the greater spreading occurred north of this zone. Then the spreading centre shifted to the Norfolk Basin (Figs. 2.27, 2.28) and finally the eastern part of North Island has slipped southwards to its present position (Fig. 2.29). During these events, geometrical stability was probably maintained by back-arc spreading, although this has not been quantitatively tested.

In the Miocene a further change in the regional spreading pattern occurred. Chase (1971) suggests that about 20 MY ago the Melanesian Border Plateau - a complex of Pacific-type volcanic islands - may have collided with the northern side of the Fiji Plateau, possibly along the Vitiaz Trench line (Figs. 2.27, 2.28). The effect of this was then to renew activity on the New Hebrides arc, but with a reversal of polarity so that oceanic crust is now consumed from the west (Fig. 2.28). Thus I have schematically suggested that the plate boundary was originally along the New Hebrides Trench, and then it shifted out to the Vitiaz Trench. Collision with the Melanesian Border Plateau then "sealed" this trench and reactivated the New Hebrides Trench with a reversed polarity. The geology of the New Hebrides appears to support this suggestion (c.f: Mitchell & Warden, 1971). A similar event may have occurred in the Solomon Islands (Figs. 2.27 - 2.29), where the trench is now south of the islands and dips steeply

north, but geological and morphological features suggesting an earlier trench-arc complex are also found on the north side of the Solomon Islands.

The most recent stages have been even more complex, and will only be mentioned briefly. The Fiji Plateau cannot decrease in size, and it is being carried towards New Caledonia, with which it will soon collide (Figs. 2.29, 2.30). The Solomon Islands are now on the Pacific plate and are shifting westwards with it. Krause (in press) and Johnson & Molnar (1972) analysed present plate motions just west of this region, thus linking the model outlined here with the Indonesian region. The Tonga/Kermadec Trench is still migrating east, and the Lau Basin has opened behind the arc within the last 10 MY (Fig. 2.30). JOIDES (1972) hole 203 indicates a Late Miocene age for the oceanic crust in this region. Fiji exhibits a complex history related to the northern end of this trench system. The "spiral" form of the bathymetry (Fig. 2.1.) suggests that some rotation has occurred.

Currently it would appear that swallowing in the Tonga Trench is proceeding so rapidly that the Tonga Arc is being forced eastwards, with accompanying back-arc spreading. The dip of the Benioff zone shallows to the north (c.f: Sykes, 1966, figs. 8, 9), where the whole of the subducted slab is under compression right down to 700 km (Isacks & Molnar, 1969). The southern end of this back-arc spreading zone (Fig. 2.30) extends into New Zealand as the Taupo Volcanic Zone (Karig, 1970). As the trench line has swung round to the east, it appears that the spreading system within the Australian plate north of New Zealand has diminished

and a new set of faults has developed to link the Alpine Fault with the Tonga Kermadec Trench (see Chapter 3.3). This present plate geometry is unlikely to last any longer than the preceding phases, and hence in the future the geometry may become even more complex.

CHAPTER 3

THE TECTONIC EVOLUTION OF NEW ZEALAND

In this chapter I shall present a synthesis of New Zealand geology based on the Gondwanaland reconstruction and subsequent movement history established in Chapter 2. This interpretation is based largely on study and general discussion of geological literature pertaining to New Zealand. The synthesis is centred around more recent references, but I would point out that much of the earlier geological literature embodies the essential information contained in later and more comprehensive works. I have not attempted to review all the literature, and in general I have referred to later work, in which references to earlier literature are contained. The aim is to present a synthesis of the gross aspects of the geological evolution of New Zealand, as a basis and stimulus for more detailed discussion in the future.

Two broad divisions can be conveniently made in the discussion:-

- (i) Paleozoic to Mesozoic evolution - within the framework of the Gondwanaland reconstruction proposed above (Section 3.2).
- (ii) Cenozoic evolution - within the framework of the pattern of continental drift established (Section 3.3).

By its very nature any objectively based tectonic synthesis has a predictive element (c.f: Appendix A for general discussion) and will thus point to many new problems which can be tackled,

both to test its validity and to further develop the synthesis by incorporation of new data. This will be apparent at many stages in the ensuing discussion. One of the major conclusions of such a synthesis is therefore to outline future work, and various possibilities for future work are mentioned in this Chapter.

3.1: A BRIEF OUTLINE OF NEW ZEALAND GEOLOGY

The first comprehensive geological account of New Zealand was published in German by Ferdinand von Hochstetter in 1864 (translated by Fleming, 1959), and outlines many of the basic geological elements. The New Zealand Geological Survey was established in 1865, and has since been the major geological investigator. James Hector, its first director, published several geological maps of New Zealand between 1869 and 1884 (c.f: Grindley *et al.*, 1959, pp. 2-5). Park (1910) and Marshall (1912) both wrote accounts of New Zealand geology with accompanying maps, and by this stage the major stratigraphic elements were fairly well defined, though the metamorphic rocks were still very poorly understood.

In 1947 the New Zealand Geological Survey published two 16 mile to the inch maps covering New Zealand, and in the following year notes to accompany these maps (N.Z. Geological Survey, 1948). On the South Island map, the Alpine Fault - a feature first named by Henderson (1937) and later recognised as extending from Milford Sound northwards to Cook Strait (Wellman and Willett, 1942) - clearly

shows up as a major lineament, though not shown as a fault at this stage. Wellman (1956) discussed the structural outline of New Zealand, and reiterated his earlier conclusion (Wellman, *in* Benson, 1952) that 480 km dextral transcurrent movement had occurred on the Alpine Fault, thus establishing it as the major structural feature of New Zealand.

In 1959 the 1:2 000 000 map and accompanying text were published (Grindley *et al.*, 1959), and these cover most of the basic elements of the geology. Subsequently the 4-mile to the inch sheets have been issued (N.Z. Geological Survey, 1959-68) and each has a legend describing the area covered. Together these publications provide a very comprehensive geological outline, which I have used as the basis for my synthesis of New Zealand. More recent work by many authors has elucidated much of the structural and metamorphic sequence in the schists and greywackes, and has added considerable detail to the understanding of many other parts of New Zealand.

Many authors have speculated on the tectonic history of New Zealand. Early ideas are reviewed by Kingma (1959), who also presents his own synthesis. More recently the concept of continental drift has received much attention, and many new attempts at tectonic synthesis have been made. I shall comment on some of these later.

The major geological elements of New Zealand are shown in Fig. 3.1, as a geological sketch-map, and in Fig. 3.2, as a generalised summary of the stratigraphic column.

Several broad divisions can be recognised:-

- (a) The (?) Precambrian to Paleozoic of western South Island, comprising sediments, schists, gneisses and intrusives.
- (b) The (?) Carboniferous to Permian Te Anau Group, of sediments, volcanics and ultramafics.
- (c) The New Zealand Geosyncline, comprising the Hokonui facies (predominantly shallow water sediments with volcanic detritus), and the Torlesse facies (greywackes, shales, and the temporally equivalent Haast Schists).
- (d) Numerous small occurrences of Mesozoic volcanics and ultramafics.
- (e) Cretaceous sediments in the East Coast and Marlborough Basins, in Northland, and at several localities in South Island.
- (f) Cenozoic sediments, widely distributed but most extensive in North Island.
- (g) Tertiary volcanics, mainly in North and South Auckland.
- (h) The Quaternary Taupo Volcanics.

Structurally the major feature is the Alpine Fault, across which several of the above divisions have been displaced. Sediments in the New Zealand Geosyncline were uplifted and deformed during the Upper Mesozoic Rangitata Orogeny. Later deformation in the Cenozoic (the Kaikoura Movements) has profoundly influenced the present outline of New Zealand. In the Quaternary large volumes of extrusives have infilled the tensional Taupo Volcanic Zone. Frequent earthquakes and Recent fault scarps indicate continuing tectonic activity.

This extremely condensed summary of New Zealand geology is far from complete, but nevertheless outlines the major features which will be discussed in the rest of this Chapter. I will endeavour to present the synthesis in chronological sequence, referring to the literature as necessary to support the interpretations and conclusions reached. Maps and diagrams, with their accompanying legends, are used to illustrate the synthesis, and the text is used for further description and discussion of these.

3.2: THE PALEOZOIC AND MESOZOIC EVOLUTION OF NEW ZEALAND

Dewey (1969b) and Dewey and Bird (1970a) have proposed a generalised evolutionary model in which a thick "geosynclinal" pile of sediments accumulates at an aseismic "Atlantic-type" continental margin; subsequently this margin becomes an active plate margin and consumption of the oceanic lithospheric plate begins; an orthotectonic orogenic then develops above the active subduction (Benioff) zone.

This basic model has been used in the interpretation of several orogenic belts (e.g. the Caledonian/Appalachian belt; Dewey, 1969a; Bird and Dewey, 1970: California; Hamilton, 1969: the Urals; Hamilton, 1970: the Andes; Rutland, 1971: etc). I have previously attempted (Griffiths, 1973) to use a similar approach to interpret the Paleozoic and Mesozoic of New Zealand. Parts of the following synthesis follow the lines of this earlier paper, and I will not refer to it frequently again. Many of the

earlier comments have been amplified, and some new ideas incorporated into the present synthesis, and in addition I have tried to evaluate it by using time-space plots.

The geological data on which this synthesis is based is taken largely from maps and publications of the New Zealand Geological Survey, which are referred to as appropriate for specific information, but have also provided much general background. Fleming (1970) gave a comprehensive summary of the Mesozoic of New Zealand, around which many of the early ideas leading to this synthesis were formulated. Landis and Bishop (1972) have also recently attempted a plate-tectonic synthesis of South Island, and while I disagree with the tectonic model which they propose (see below) their paper amplifies and generally agrees with much of the stratigraphic, structural and metamorphic summary presented here, and includes many references to otherwise unpublished data.

I have already presented evidence which indicates that New Zealand remained part of Gondwanaland until the Upper Cretaceous (Chapter 2.4), and indicated the general configuration of New Zealand before drift occurred (Chapter 2.2). In Fig. 3.3 (based on Griffiths, 1973, fig.2) the major structural elements of New Zealand at the present (left - from Fig. 3.1) are shown superimposed on the Mesozoic reconstruction (right - from Fig. 2.8). Thus the paleogeographic framework for the Paleozoic and Mesozoic synthesis is established.

Paleogeographic interpretations of New Zealand now appear much more simple than previously indicated (c.f: Fleming, 1962, Figs. 3, 4;

Quennell, 1964, Fig. 2; in these figures the initial "geosyncline" is shown as a sinuous belt) and indicate a general eastward progression from foreland to ocean floor, across a continental margin. Fleming (1970) has summarised the geology of this sequence, from Paleozoic foreland, passing eastwards through continental shelf sediments (Hokonui facies) to continental slope greywackes (Torlesse facies). The Hokonui and Torlesse facies together constitute the New Zealand Geosyncline, which has been interpreted as a sedimentary accumulation at a continental margin (Landis and Coombs, 1967; Fleming, 1970; Griffiths, 1973). [In this discussion I will continue to use the established term "New Zealand Geosyncline" as a convenient reference to the Permian-Jurassic sediments, without implying any genetic or tectonic significance].

3.2.1: THE PALEOZOIC FORELAND

The foreland of the New Zealand Geosyncline is composed of (?)Precambrian and Paleozoic rocks. No definite Precambrian is known in New Zealand, though sediments of the Greenland and Waiuta Groups have been thought to be Precambrian on the basis of their lack of fossils (Grindley *et al.*, 1959 p.13). However, Laird (pers. comm.) speculates that the Greenland Group at least may be Lower Paleozoic. Aronson (1968) determined Pb^{207}/Pb^{206} ages ranging from 1170 to 1480 MY on rounded detrital zircons from the Greenland Group, and from this concluded that an acidic continental terrain must have existed close to New Zealand to supply these zircons. Solomon and Griffiths (1972) argue that a large Precambrian block stretched from Tasmania to New Zealand,

based on a tectonic analysis of the Tasman Orogenic Zone (see Section 2.3.1), and this block may thus be the acidic terrain suggested by Aronson (1968).

Paleozoic (pre-Te Anau Group) rocks occur in Nelson, Westland and Fiordland (Fig. 3.1). The most extensive outcrops are west of the Alpine Fault in Nelson and north Westland, and comprise a varied Cambrian to Lower Devonian sequence of sediments, volcanics, ultramafics, schists, gneisses and intrusive granitic rocks (Grindley, 1961a). In Fiordland, Ordovician sediments occur at Chalky and Preservation Inlets, and much of the regional metamorphic terrain is also considered to be Paleozoic (Wood, 1960).

Aronson (1968) obtained ages ranging from 285 to 365 MY from several igneous and metamorphic rocks from Nelson and Westland, and concluded that a climax of metamorphic activity occurred about 350-370 MY ago. Some of his age determinations also indicate a later Cretaceous "overprint", and hence it seems quite probable that other rocks which now give apparent Cretaceous ages may in fact be Paleozoic. Turner (1938a, p.237) recognised that gneisses in the Lake Manapouri region had been reconstituted during a later phase of regional metamorphism *prior* to intrusion of granites, and since the granites and metamorphism are of Cretaceous age (c.f: Aronson, 1968), it is probable that the original gneisses were Paleozoic or older.

The Paleozoic rocks exhibit similarities to rocks in eastern Australia, and may thus be considered part of the Tasman Orogenic Zone (Solomon and Griffiths, 1972). Occurrence of Devonian granites also indicates a Late Paleozoic cratonisation of the crust extending

as far east as New Zealand, which then formed the "foreland" for later events.

The Te Anau System (now Te Anau Group) was defined by Grindley (1958, p.21) to include "all the post-Devonian sediments, volcanics and intrusives deposited or erupted prior to the deposition of the Lower Permian Maitai limestone". Te Anau Group rocks occur in Nelson, Southland and Otago (Fig. 3.1), and were a continuous belt before being disrupted by the Alpine Fault (Wellman, 1956). The Te Anau Group consists of many formations, discussed in detail by Grindley (1958, pp.21-40). Volcanism was important throughout this phase in the geological evolution, and I would interpret the Te Anau Group as being closely related to an Upper Carboniferous to Lower Permian subduction zone along the foreland margin. I will not elaborate this in detail, but there is evidence to support the hypothesis:-

- (a) The variety, sequence and associations of rock-types are similar to those related to other subduction zones — i.e: sediments (argillites, green sandstones, spilitic tuffs, volcanic breccias, etc.), volcanics (spilites, keratophyres, gabbroic intrusives, etc.), and ophiolites (norites, dunites, peridotites, pyroxenites, serpentinites, etc.) all in intimate association (c.f: Grindley, 1958, pp. 21-40).
- (b) The overlying Hokonui Facies contain abundant volcanic detritus, which may in part have been derived from erosion of an earlier volcanic belt, and in part reflect continuing volcanic activity on the western foreland.

(c) Challis (1968) shows that the $\text{Na}_2\text{O}:\text{K}_2\text{O}$ ratio in the Te Anau Group volcanics increases to the ~~west~~^{east}, and thus away from the postulated subduction zone, as has been determined in other volcanic arcs which are interpreted as having formed above Benioff zones.

Thus I envisage the Te Anau Group volcanics, ophiolites, etc., to have formed as a result of subduction at an essentially "Andean-type" coast (Fig. 3.4). Further investigations could be carried out to test this hypothesis, such as geochemical analyses of the volcanics.

It is of historical interest to note that over a hundred years ago, Hector (see Grindley *et al.*, 1959, p.17), seeing the Te Anau beds in Greenstone Valley, Southland, was reminded of "Darwin's great porphyritic formation" in the South American Andes, probably the first suggestion of their relation to an Andean-type coast.

3.2.2: DEPOSITIONAL HISTORY OF THE NEW ZEALAND GEOSYNCLINE

The depositional history of the New Zealand Geosyncline can be broadly discussed in terms of two units - the Hokonui and Torlesse facies (see Fig. 3.5). The contrast between these two units has long been recognised. The Hokonui facies [also referred to as "western" or "marginal" facies] were defined by Wellman (1952) and are considered to have been deposited on a continental shelf (c.f: Fleming, 1970). The Torlesse facies [also referred to as "Alpine", "eastern" or "axial" facies] were defined by Fleming (1970) to include both "greywackes" (Torlesse Group of Suggate, 1961) and the Haast Schist Group (Suggate, 1961) which are temporal equivalents,

and are interpreted as having been deposited on a continental slope, perhaps largely on oceanic crust (c.f: Landis and Coombs, 1967; Fleming, 1970).

Deposition of the Hokonui facies began in the Lower Permian. In Southland, limestones (Maitai Series) either rest conformably on the Te Anau Group, or above intervening volcanic breccias (Grindley, 1958, p.42). A fossiliferous sequence of predominantly shallow water clastics, with abundant volcanic detritus, and occasional limestones, was then deposited fairly continuously until the Jurassic. In Southland the Hokonui facies appear to thicken markedly eastwards (Mutch, 1957, fig.2), as might be expected on a continental shelf. The Hokonui facies are now preserved in three major synclinoria (Fig. 3.3), and hence may originally have covered a larger area (Fig. 3.5).

A major problem in the Hokonui facies which is critical in understanding the tectonic history, is the origin of the abundant volcanic detritus in the sediments (e.g: Coombs, 1954; Wood, 1956; Mutch, 1957; Fleming, 1970; Dickinson, 1971). These authors have explicitly or implicitly assumed that the occurrence of volcanic detritus is indicative of *contemporaneous* volcanism, presumed to be on the foreland to the west, throughout the Permian to Jurassic. Dickinson (1970, 1971) and Landis and Bishop (1972) attempt to reconcile this conclusion with a plate tectonic model, in which volcanism is related to contemporaneous subduction, and hence propose that subduction, with an associated trench and Benioff zone, was continuous throughout most of the Permian to Jurassic. Thus they imply that the Torlesse facies were largely deposited near to or in an active oceanic trench.

However, I see considerable difficulties inherent in this model, which implies continuous "scraping off" of the Torlesse facies from the subducting oceanic plate and severe deformation contemporaneous with deposition. Much of the Torlesse facies (the greywackes) are unmetamorphosed, and do not display the totally chaotic deformation which would be expected if they were within an active trench. I would instead interpret the Hokonui and Torlesse facies together as being deposited at an *aseismic* ("Atlantic-type") continental margin, with no subduction occurring between the Permian episode and the Mid-Jurassic.

This interpretation raises the question of the source of the volcanoclastic material in the Hokonui facies, and I would make the following suggestions:-

- (a) Some of the volcanic fraction may be the product of erosion of an *older* volcanic arc, rather than reflect contemporaneous volcanism. Most petrographic descriptions (e.g: Wood, 1956) emphasise the reworked nature of the volcanic material, even though much of it is still fairly fresh-looking - implying that it was derived from a nearby source. The work of Dickinson (1971), based on sedimentological studies, does not appear to contradict this view, as he refers several times to the "reworked" nature of much of the volcanic detritus. Criteria for distinguishing between contemporaneous and earlier volcanic activity purely from detrital volcanic debris may be very subtle (Dr. K. Crook, pers. comm.), and thus a very careful study of the Hokonui sediments may be needed.

- (b) Vitric material is most abundant in the Permian and Lower Triassic. This material may have been derived from continuing volcanism in the western foreland, perhaps in part related to injection of granites at depth some distance from the former trench to the east. Aronson (1968, p.682) gives a date of 212 MY on a muscovite from a granite within the Charleston Gneiss in Nelson. This granite lies between older Devonian granites to the west, and younger Cretaceous granites to the east (Fig. 3.1), which is consistent with my model of episodic eastward-migrating subduction. More age determinations would be valuable to try to identify other Permo-Triassic intrusives in this area. Volcanic activity may thus have persisted up to 50 MY or more after subduction ceased in the Lower Permian, and there is no justification in assuming continuous subduction during this period.
- (c) The general trend in composition of the clastic igneous fraction is an upward increase in the ratio of plutonic to volcanic material (Coombs, 1950; Wood, 1956), which is consistent with progressive downward erosion of an earlier volcanic terrain developed on continental crust. I have suggested above that such a volcanic arc may be represented by the Upper Carboniferous to Permian Te Anau Group. Challis (1968) recognised two distinct belts of basic igneous rocks - one Permian and one Cretaceous.

- (d) There is much more positive evidence for *renewed* subduction with a new phase of volcanism in the Mid-Jurassic (see Section 3.2.3). If subduction had been continuous from Permian to Cretaceous, a great deal of evidence for the actual volcanics might reasonably be expected, whereas none has been found.

Though I have referred to the Permian to Jurassic margin as "Atlantic-type", I am not implying that it was a pull-apart margin, but only that it was aseismic (i.e.: not a plate margin) over this period. The contrast between the Hokonui Facies, with their volcanic detritus, and the "miogeosynclinal" sediments of pull-apart margins, reflects therefore the different tectonic setting in which the sedimentary pile developed -- in New Zealand there never was an eastern continental block.

Determination of the source of the volcanoclastic fraction of the Hokonui facies is thus important in understanding the tectonic evolution of New Zealand during the Mesozoic. The interpretation which I suggest is, I consider, consistent with the geological data. It differs from others (e.g.: Dickinson, 1970, 1971; Landis and Bishop, 1972) principally in the chronology of events early in the evolution of the New Zealand Geosyncline, and in so doing it corresponds much more closely than the others with the scheme of orogenesis proposed by Dewey (1969b). This conflict in interpretation is as yet unresolved, but thus poses a problem which, in its solution, will contribute both to further understanding of the geology of New Zealand, and to the general theory of plate tectonics.

The Torlesse facies are the deeper water equivalent of the Hokonui facies, and comprise a very thick, monotonous sequence of greywackes and argillites, with redeposited turbidites, and, locally, spilites and cherts (Fleming, 1970). The oldest sediments recorded are Carboniferous (Jenkins and Jenkins, 1971), thus implying that deposition had begun at least locally at the continental margin by this time. The Torlesse facies greywackes pass into, and are therefore the equivalent of, the Haast Schist Group (Suggate, 1961).

Landis and Bishop (1972) discuss the Torlesse facies in much more detail. They conclude that a western landmass was the source of both Torlesse and Hokonui facies, but point to differences in the two facies which they consider implies that each was deposited in a discrete basin. They develop a plate tectonic model to attempt to explain the apparent general westward younging of the Torlesse greywackes, which involves an active trench to the east throughout the period of deposition. However, as I have previously argued, the presence of volcanoclastic material in the Hokonui facies is not conclusive evidence for continuous subduction, nor do the Torlesse facies exhibit the characteristics which might be expected if they had been deposited in or near to an active trench. The apparent westward younging could equally well be related to differential uplift and erosion during the later Rangitata Orogeny or Kaikoura Movements.

I consider that the overall relationships are much more convincingly interpreted in terms of sedimentary accumulation at an aseismic "Atlantic-type" continental margin from Permian to Mid-

Jurassic. Thus, following the Lower Permian episode of subduction, the Pacific margin of Gondwanaland in New Zealand became a stable continental margin, at which a thick wedge of clastic sediments, derived from the craton to the west, was deposited (see Figs. 3.5, 3.6a). Final vestiges of volcanism continued into the Triassic, thus providing a volcanoclastic fraction in the otherwise normal shelf-type Hokonui Facies.

The Hokonui and Torlesse facies are now separated by an abrupt tectonic break (c.f: Fleming, 1970; Landis and Bishop, 1972). This break is associated with Tertiary faults (Fleming, 1970), and in Southland Kingma (1959) interpreted it as the line along which Torlesse facies had been thrust westwards over Hokonui facies. Such an interpretation is consistent with the model for the evolution of the New Zealand Tertiary based on the sea-floor spreading history of the south-west Pacific which I shall later outline (Section 3.3). Thus the now apparent tectonic break may not have existed prior to the Tertiary, and the Hokonui and Torlesse facies have interfingered as I have indicated (Fig. 3.6a), giving a cross-section similar to that of present-day "Atlantic-type" continental margins. Campbell and Warren (1965) record shell beds, plants and conglomerates, reflecting shelf deposition, within the Torlesse facies, which suggests that at least in some places the shelf conditions of the Hokonui extended well out into the Torlesse facies without any break during some periods. Thus the simple model which I have proposed to explain the depositional history of the New Zealand Geosyncline appears to be consistent with geological data, and in addition has resulted in predictive interpretations which are capable of being

used to more thoroughly evaluate the synthesis.

3.2.3: *THE RANGITATA OROGENY*

The earliest tectonic movements associated with the Rangitata Orogeny began about Mid-Jurassic, after which uplift of the New Zealand Geosyncline resulted in progressive cessation of sedimentation (Wellman, 1956; Fleming, 1970). Many of the subsequent events can be interpreted in terms of the model proposed by Dewey (1969b) for the evolution of an orthotectonic orogen developing at a continental margin. I have schematically illustrated the model by a series of cross-sections (Fig. 3.6, modified from Griffiths, 1973, fig.4), and paleotectonic maps (Figs. 3.7 to 3.9).

A new Benioff zone thus probably developed in the Mid-Jurassic, and consumption of the Pacific (oceanic) lithospheric plate commenced (Figs. 3.6, 3.7). It is difficult to precisely date the onset of subduction. Fleming (1970) indicates that the first orogenic movements occurred in the Middle Jurassic, based on the observation of Speden (1961) that there is evidence of eastward thinning of some of the Jurassic Hokonui formations in Southland. Fleming (1970) also points out that widespread non-marine phases occur in the Middle Jurassic in Southland, but that deposition of marine facies continued in the north (e.g. in the Kawhia Syncline - see Kear, 1960). It appears therefore that subduction began about early Middle Jurassic, and that it may have been a diachronous event. It must have pre-dated the metamorphism in the Otago Schists (about 140 MY ago - Harper and Landis, 1967)

in South Island, and the occurrence of Lower Cretaceous volcanics (see below), though many of the latter have not been accurately dated. These events taken together suggest a tentative date of about 160 MY for the beginning of subduction.

There is considerable evidence for renewed volcanism associated with this subduction (Figs. 3.7, 3.8), though many of the volcanics do not appear to have been studied in detail, particularly geochemically. Volcanic activity becomes prominent in the Upper Jurassic - Lower Cretaceous (Wellman, 1956). A variety of igneous rocks occur, and they are similar to those found associated with other subduction zones (c.f: Dewey and Bird, 1970a, 1971). These are indicated in Fig. 3.1, and include:-

- (a) Matakaoa Volcanics - basic intrusives, pillow lavas and tuffs, in association with greywackes, chert and some limestone, of Upper Jurassic age (Grindley *et al.*, 1959, p.49; Kingma, 1965).
- (b) Tangihua Volcanics - similar to (c), of probable Upper Jurassic age (Grindley *et al.*, 1959, p.47-8; Thompson, 1961; Kear and Hay, 1961; Hay, 1960, pp. 49-55).
- (c) North Cape Ultramafics - ultramafics and gabbros, with serpentinites, representative of the ophiolite suite (Grindley *et al.*, 1959, p.47; Kear and Hay, 1961).
- (d) Mount Camel Volcanics - keratophyres and spilites, associated with greywacke and argillite, probably Jurassic (Grindley *et al.*, 1959, pp. 46-7; Kear and Hay, 1961).
- (e) Mount Somers Volcanics - rhyolites, andesites and dacites, originally thought to be Upper Jurassic or Lower Cretaceous

(Speight, 1938; Grindley *et al.*, 1959, p.46; Gair, 1967).

Hulston and McCabe (1972 p.419) however correlate the Mount Somers Volcanics with andesite, from a drill hole near Christchurch, which has an age of 81 MY.

- (f) Intrusive syenites and gabbros in the Mandamus district (Mason, 1951); pyroxenite, syenite and gabbro intrusions in the Inland Kaikoura Range and alkali-olivine basalts in the Clarence and Awatere Valleys (Mason, 1958). These all intrude the Torlesse Facies, and are probably of Lower Cretaceous age.

Taken together (a), (b), (c) and (d) are interpreted as a trench assemblage. On the reconstruction (Fig. 3.3) the whole of the Northland sequence (b,c,d) lies well to the west of the postulated trench zone. However, Wellman (pers. comm.) and Temple (1972) indicate that much of this complex may be allochthonous, and was derived from the east in the Tertiary (see also Kear and Waterhouse, 1967). Isostatic gravity anomalies (Reilly, *in* Thompson and Kermode, 1965) suggest that the volcanics and ultramafics are thin sheets with no "roots", and their contact with the underlying rocks are usually mapped as faults (Thompson, 1961; Kear and Hay, 1961). This allochthon was probably emplaced in the Miocene (see Section 3.3.2). The other groups (e,f) both intruded the Torlesse Facies during the Cretaceous, post-dating the major deformation phase of the Rangitata Orogeny, and possibly up to 50 MY after the cessation of subduction (see Section 3.2.4). [This time relationship is similar to that previously suggested for that between volcanism and

subduction in the Permo-Triassic]. Challis (1968) includes all of these groups in a "Cretaceous volcanic belt", and shows that there is an increase from west to east in the $\text{Na}_2\text{O} : \text{K}_2\text{O}$ ratio similar to that found passing from the continental to the oceanic side of present-day volcanic arcs.

There is thus good evidence for a definite episode of volcanism beginning in the Upper Jurassic and continuing to the Lower Cretaceous, which can be explained in terms of subduction along the Pacific margin during the Rangitata Orogeny. This is in marked contrast to the continuous volcanism which should be evident if subduction had been a Permian to Jurassic event. Further study of all these volcanics seems warranted in view of their tectonic significance. It is also worth noting that Dewey (1969b) sees calc-alkaline volcanic activity as occurring *after* the peak of deformation and metamorphism, and in New Zealand these volcanics (e, f, above) are resting on already deformed greywackes.

The Rangitata Orogeny is thus interpreted as being a direct consequence of subduction at the Pacific margin beginning about Mid-Jurassic. Subsequently the New Zealand Geosyncline became an orthotectonic orogen in which the sequence and style of events are comparable to those in the plate tectonic model proposed by Dewey (1969b) for orogenesis at an Andean-type margin.

The sediments of the Torlesse facies were uplifted, and underwent complex, polyphase, deformation and metamorphism during the orogeny. As now exposed, there is a general westerly trend towards increasing metamorphism and decreasing age (Landis and Bishop, 1972). This probably results from (a) the relative amount of contem-

poraneous uplift and variation of geothermal gradient across the orogen, and (b) superimposed and unrelated effects of Tertiary uplift resulting from sea-floor spreading (see Section 3.3). The least effects of (b) are probably in eastern Otago and Southland, well away from the Alpine Fault. Here the most intense belt of deformation seems to be flanked by less intense belts (Landis and Bishop, 1972), possibly reflecting the relative amounts of uplift across the orogen. Further studies may help to elucidate in detail the thermal regime which existed across the orogen as it was developing.

Landis and Bishop (1972) discuss the metamorphic zonation in the Torlesse facies in detail. The main zones which they recognise (see Fig. 3.1) are as follows:-

- (i) an eastern terrain of non-foliated, greywacke-type sediments, of zeolite-prehnite-pumpellyite facies, grading into
- (ii) semi-schists of pumpellyite-actinolite facies, grading into
- (iii) greenschist (chlorite-biotite zone) facies schists (the Otago Schists).
- (iv) Higher grades (amphibolite facies) occurring near to the Alpine Fault in the Southern Alps.
- (v) In Otago the grade of metamorphism decreases southwards into pumpellyite-actinolite schists and prehnite-pumpellyite metagreywackes, and locally lawsonite-albite-chlorite metagreywackes occur.

Overall these assemblages indicate that a relatively high level of the orogen is exposed. Several stages of deformation and metamorphism have occurred in the Torlesse facies (e.g: Turner, 1940; Brown, 1963, 1968; Grindley, 1963; Wood, 1963; Means, 1966). There is a striking similarity to the sequence elucidated in the Scottish Highlands (e.g: Johnson, 1963), which has been interpreted as an orthotectonic orogenic belt by Dewey (1969a,b).

Detailed mapping in New Zealand has revealed large nappes and recumbent folds (e.g: Means, 1966; Brown, 1968; Bishop, pers. comm.), and hence the present age distribution of Torlesse facies rocks (c.f: Landis and Bishop, 1972, fig.2) may not necessarily reflect any original zonation across the belt. Considerably more detailed work remains to be done to fully elucidate the stratigraphic-structural-metamorphic patterns developed. Reference to tectonically similar terrains elsewhere (e.g: Scotland) may thus at least provide ideas as to the critical relationships which should be sought in the field.

The Hokonui facies exhibit a much simpler pattern of deformation, probably because they are a thinner unit resting essentially on continental crust some distance from the trench. It is necessary to distinguish between Tertiary and Mesozoic (Rangitata) deformation (see Section 3.3), and again Otago/Southland may have suffered least during the Tertiary (c.f: Fleming, 1970, Fig.3). Metamorphism is generally of the prehnite-pumpellyite, lawsonite-albite-chlorite, and zeolite facies (Coombs *et al.*, 1959; Landis and Coombs, 1967). The rocks of the Te Anau Group are similarly metamorphosed, and lawsonite-bearing glaucophane schists occur locally in

the tight axial regions of synclines, probably as a result of later "tectonic overpressure" (Landis and Coombs, 1967) during the Tertiary deformation.

Overall, the New Zealand Geosyncline and its foreland became a "paired metamorphic belt" (Miyashiro, 1961) during the Rangitata Orogeny (Fig. 3.8; Landis and Coombs, 1967). The metamorphic facies developed in the New Zealand Geosyncline characteristically lie on low to intermediate T/P gradients (Turner, 1968, Fig. 8-4), whereas the foreland province was intruded by granites during the Rangitata Orogeny, indicating a high T/P gradient. Retrogressive metamorphism of high-grade schists and gneisses has also been noted (e.g: Turner, 1938a, p. 237; 1938b, pp. 172-3; Hutton, 1940, p. 73). Thermal effects have presumably also caused argon loss from the older gneisses, which now yield Cretaceous ages (Aronson, 1968). The overprint of a high grade regional metamorphic assemblage by a later thermal event might also be expected to provide interesting material for petrological studies.

The culmination of metamorphism in the low T/P belt appears to have slightly post-dated the major structural deformation (c.f: Table 3.1), and radiometric dating (see Fleming, 1970, Fig. 18) also suggests that metamorphism in the high T/P belt (100-120 MY) may have occurred slightly later than in the schists (130-140 MY). A similar age zonation has been noted in other orthotectonic belts (c.f; Dewey and Pankhurst, 1970).

3.2.4: *EVENTS CONSEQUENTIAL TO THE RANGITATA OROGENY*

During the Rangitata Orogeny, deformation and uplift of

the New Zealand Geosyncline formed new mountains (here referred to as the Rangitata Mountains) which were themselves eroded and thus provided a new source of sediments. Syn- and post-orogenic "molasse" and "flysch" sediments were deposited respectively on the western and eastern sides of the emergent mountains [I use these terms in the same sense as Dewey, 1969b, i.e: flysch = sediments derived by erosion of a rising orhotectonic orogen and deposited on the oceanic side; molasse = sediments stripped from the orogen and deposited in troughs in the foreland]. The most extensive exposures of these rocks are in North Island, reflecting the greater Tertiary uplift and erosion in South Island (see Section 3.3). The Cretaceous paleogeography, as did the earlier Mesozoic, becomes considerably simplified when plotted on the reconstruction. Lower Cretaceous (Taitai Series) and Upper Cretaceous (Clarence, Raukumara and Mata Series) paleogeography is shown in Figs. 3.8 and 3.9 respectively.

The oldest syn-orogenic flysch deposits recognised belong to the Taitai Series, and occur mainly on the East Coast of North Island (Wellman, 1959; Grindley *et al.*, 1959). In the East Cape area (Kingma, 1965) the Taitai Series are a sequence of alternating conglomerates, grits, sands and silts. The conglomerates contain pebbles of assorted acidic to basic volcanics, acidic plutonics and cherts (Ongley and MacPherson, 1928; Kingma, 1965), probably derived from contemporaneous igneous activity to the west.

The Taitai Series are thus early flysch derived from the rising orogen, the sediments being a mixture of reworked greywacke from the Torlesse facies, and igneous debris from the volcanics and

basalts related to the subduction zone (Fig. 3.8). The apparent eastern source of some material (Kingma, 1965) may thus reflect some uplift of the outer edge of the continental margin near to the trench. The Taitai Series of East Cape range upwards from about middle Lower Cretaceous (about 110-120 MY ago), and may thus give a rough estimate of the time at which subduction was ending, as overlying beds reflect quieter tectonic conditions. Wellman (1956) and Fleming (1970) both conclude that a rapid sequence of events took place in the Lower Cretaceous, with rapid emergence of the New Zealand Geosyncline and deposition of much eroded and reworked greywacke material, and that tectonic activity became insignificant in the Upper Cretaceous. Thus I would tentatively estimate that Mesozoic subduction at the Pacific margin occurred over a period of no longer than about 50 MY (i.e. Middle Jurassic to Lower Cretaceous), and this was sufficient to produce all the structural and thermal events associated with the Rangitata Orogeny. In terms of the order of magnitude of the total amount of Pacific oceanic crust consumed, this would be, over a period of 50 MY, 2500 km at a swallowing rate of 5 cm/yr, or 5000 km at 10 cm/yr.

Taitai Series rocks also occur in the Wairarapa district, where they are virtually indistinguishable from underlying Upper Jurassic beds containing spilites (Kingma, 1967) which may indicate an active trench to the east during this period. Evidence of an eastern source can again be interpreted in terms of uplift near to the trench, rather than as evidence for the "eastern cratonic ridge" advocated by Kingma (1959). In Marlborough, Taitai Series rocks are included in the Upper Torlesse facies by Lensen (1962),

who comments that the presence of conglomerates with boulders of greywacke indicates penecontemporaneous erosion of older geosynclinal deposits was occurring.

Mid-Cretaceous sandstones and conglomerates also occur in the Chatham Islands (Hay *et al.*, 1970) and in New Caledonia (Lillie and Brothers, 1970), suggesting that flysch sedimentation was continuous along the Pacific margin following cessation of subduction. This interpretation is in contrast to earlier suggestions, which do not take account of continental drift (e.g: Grindley, 1961b, fig.3; Kingma, 1959, Fig. 16; Quennell, 1964, Figs. 3,4), that the New Zealand Cretaceous was deposited in a series of small unconnected "geosynclines" with postulated eastern landmasses which have subsequently vanished.

Above the Taitai Series the eastern flysch sediments show a gradual upward transition to shelf sediments through the rest of the Cretaceous, represented by the Clarence, Raukumara and Mata Series (Fig. 3.9; also Wellman, 1959, especially fig. 20). In the East Coast region the Clarence Series are generally similar to the underlying Taitai Series. Flysch content decreases upwards, and the Mata Series are predominately a fossiliferous shelf sequence of sandstones and calcareous silts (Kingma, 1962, 1965, 1967).

In Marlborough the Clarence Series comprises sandstones and greywacke-derived conglomerates with interbedded basalt flows (Clarence Basalts) (Suggate, 1958), which may represent a late restricted phase of igneous activity. In the Raukumara and Mata Series there is evidence for a gradual marine transgression westwards (Lensen, 1962), indicating progressive erosion of the Rangitata

mountains. Small outcrops of predominantly marine sandstones and shales of the Clarence, Raukumara and Mata Series also occur in Northland (Kear and Hay, 1961, Thompson, 1961).

Molasse occurs at several localities in South Island (Figs. 3.8, 3.9). The oldest beds occur in the Pororari Group in south Nelson and north Westland (Bowen, 1964; Mutch and McKellar, 1964; see also Suggate, 1957). This Group lies unconformably on older Paleozoic rocks and consists of non-marine conglomerates, sandstones, silts and some coals towards the top. The conglomerates contain much angular granitic debris (e.g: Hawks Crag Breccia, see Suggate, 1957). The base of the Group is possibly mid-Upper Jurassic (Norris, 1968), indicating that rapid erosion of the Rangitata mountains was occurring by this time. The Paparoa Group occurs in the Greymouth Coalfield in Westland, and is equivalent to Clarence and Raukumara Series. The sequence, which rests unconformably on Paleozoic rocks, consists of conglomerates, sandstones, shales and coals (Gage, 1952). Although these West Coast outcrops are small in area, they are important in that they indicate that similar molasse may have once covered a much larger area of the foreland. The outcrops preserved were originally some distance from the mountain belt (see Fig. 3.8), and rest on rocks which did not suffer a Cretaceous thermal overprint (see Aronson, 1968, Fig.4) as they were several hundred kilometres west of the trench.

The Kaitangata Coalfield in Otago comprises a Cretaceous freshwater sequence with greywacke-derived breccias and conglomerates, passing upwards into a marine transgressive sequence (Harrington, 1958). At several other widely separated localities there is also evidence

for rapid deposition of clastic material in the Upper Jurassic to Lower Cretaceous, until finally the Rangitata mountains had been eroded to sea-level, and a marine transgression occurred across New Zealand, continuing into the early Tertiary (c.f: Wellman, 1959; Fleming, 1970). This transgression marks the end of the Paleozoic and Mesozoic evolution of the New Zealand Geosyncline.

3.2.5: *TIME-SPACE ANALYSIS*

Finally in this section I will evaluate the evolutionary synthesis of the Paleozoic and Mesozoic of New Zealand, which I have outlined above, using time-space plots. These plots have been developed to enable an objective portrayal of geological data, in particular to show the relative positions in time and space of the various geological events of an area. The basis and method of construction of these plots, and their use in tectonic synthesis, is discussed fully in Appendix A, and only further details pertinent to the New Zealand plots will be added here.

I have discussed the Mesozoic geology of New Zealand in some detail above (Sections 3.2.2; 3.2.3; 3.2.4), and interpreted many features in terms of the plate tectonic models proposed by Dewey (1969b) and Dewey and Bird (1970a). In order to critically evaluate this synthesis using time-space plots it is necessary to construct separate plots based (a) on the available geological data, and (b) on an actualistic model, and then to test their compatibility (see Appendix A.4; A.5).

A time-space plot of New Zealand geology is presented in Figure 3.10. This plot has been constructed using data gathered from the literature referred to in the preceding sections. It represents an attempt to portray the relations of events in time and space without making any tectonic interpretations, other than that New Zealand was part of Gondwanaland during the Mesozoic (i.e: the Alpine Fault did not yet exist) as previously established (Chapter 2.5).

The time axis is to scale from the Permian upwards, and positioning of events depends mainly on paleontological and radiometric data. The space axis does not represent any single section which can be drawn through New Zealand, but a composite "stack" of geological data (see Appendix A, section 3). The inset (lower right) is an outline of New Zealand at the present day. The plus, zero and minus signs roughly indicate relative levels of erosion. The West Coast is taken as an arbitrary zero, though this has no absolute significance. Thus in comparison Southland and Otago are considerably uplifted, due to Tertiary movements on the Alpine Fault, whereas in comparison much of North Island is relatively downwarped and the deeper levels are buried beneath Cretaceous and Cenozoic sediments. Different parts of the geological column are therefore best exposed in different localities (Fig. 3.2).

The data used in constructing the plot is taken from all of New Zealand as appropriate at different times. The plot aims to show the *relative* positions in space of events, perpendicular to the foreland margin, through time. Thus events on the foreland are best seen on the West Coast and in Southland/Nelson, so the left side

of the plot is based on these areas. Similarly the lower and upper levels of the New Zealand Geosyncline are seen in Otago and Auckland, and the syn- and post-orogenic sediments on the East Coast and in small areas on the West Coast. Plotting data in this way does not give so precise a result as using a single cross section, and will not show any diachronous trends *along* the margin. A more accurate picture could be obtained by drawing several plots across the margin. However, a single "stacked" plot is a very useful way to summarise much of the geological data in a clear visual manner.

The Paleozoic and Te Anau Group (TAG) are only shown schematically on the plot. Above these the New Zealand Geosyncline shows up as the western Hokonui Facies (MS; GS; BS; HKOS) and the eastern Torlesse facies (HS; "UG"), building out eastwards onto oceanic crust. Basaltic igneous activity (TV; MV) occurs in the Upper Jurassic, and at the same time deformation of the Geosyncline begins. Non-marine clastics (PoG) are deposited on the foreland. In the Lower Cretaceous, metamorphism of two contrasting types (high T/P with associated granitic intrusives; low T/P) occurs, followed by acid volcanics (MSV). Reworked clastics are deposited on the oceanward side of the uplifted mountains (TS). In the Upper Cretaceous more clastics occur on the foreland (PG) and on the eroded orogen and a gradual transition upwards towards a marine transgression takes place (CRMS; DS). Upper Tertiary events are again schematic - first quiet marine deposition (AS; LS), and then the Alpine Fault cuts across the region in the Miocene, with accompanying diastrophism, and in the Quaternary the Taupo Volcanic Zone (TVZ) develops.

The plot therefore summarises much of New Zealand geology. I would again emphasise that the basic data is obtained from the

literature only. When extracting the data to compile the plot, the excellent quality of the geological work done in New Zealand was very apparent, as much of the literature contains clear, objective geological descriptions without tectonic speculations. Interpretations of rocks as trench assemblages, post-orogenic flysch and molasse, etc., are not made at this stage, but the nature of sediments, igneous and metamorphic rocks, etc., can be recorded symbolically.

I have proposed that the geological evolution can be synthesised by using a similar model to that of Dewey (1969b) and Dewey and Bird (1970a) for conversion of an Atlantic-type continental margin to an Andean-type margin, based on the plate tectonic hypothesis. Therefore a time-space plot of this model must now be constructed, in order to test its compatibility with the geological data. I have initially taken the time-space plot of Dewey and Bird (1970a, fig. 15) as representative of their model.

The derivation of both actual and model time-space plots is illustrated in Fig. 3.11, as a flow-chart. The actual plot, described above, is constructed using literature based on field data. The model plot is based on the plate tectonic model for the evolution of an orthotectonic orogen. The two are "compared" in the centre diamond. The plot of Dewey and Bird (1970a, fig.15) has no time-scale, and hence it is necessary to vary the overall dimensions of it until it is similar to the actual plot. This does not affect its compatibility with the plate tectonic hypothesis, rather it adds to the *model* the constraints of a *real* situation.

I conclude that an acceptable comparison exists between the

two plots, and that a synthesis of the New Zealand Mesozoic based on plate tectonics is possible. This is illustrated in Fig. 3.12. The upper plot is based on Dewey and Bird (1970a, fig.15). The lower plot is an *interpretation* of Fig. 3.10 in terms of the proposed synthesis, and should be carefully compared with Fig. 3.10. Trench assemblages (TV, MV), volcanics (MSV), molasse (PoG, PG), flysch (TS), etc., are now identified in New Zealand, and the overall similarity of the plots is obvious.

Comparison of the two plots of Fig. 3.12 also indicates some complications. Dewey and Bird (1970a, fig.15) show flysch on *both* sides of the rising orogen. In my description of New Zealand based on the model of Dewey (1969b), I followed his nomenclature of flysch (oceanward side) and molasse (foreland side). The inconsistency is not mine, as Dewey and Bird (1970a) use a different terminology. Their "foreland" flysch is probably equivalent to the continuing Upper Jurassic deposition in the Kawhia Syncline (see Section 3.2.3) and to other shelf sediments which may have been subsequently eroded away. Their model is not exactly equivalent to New Zealand in timing, as "molasse" occurs much earlier than they indicate. This is but one example of a feed-back from the actual geology exercising a constraint on the model. This is readily apparent using the scheme of analysis shown in Fig. 3.11, thus illustrating the advantages of a logical approach to tectonic analysis.

Fig. 3.11 also shows the relationships of "feed-back" and "predictive" tests to the synthesis (compare Fig.A.1). In this case the synthesis allows a feed-back to the geological literature,

which should contain extra data compatible with the synthesis, or suggest new interpretations, and also allows feed-through constraints to be applied to the model as suggested above. The synthesis of New Zealand points to many such areas, some of which I have mentioned whilst discussing the geology. These are perhaps the most important contribution made by the synthesis.

Having established that the plate-tectonic hypothesis provides the basis for a satisfactory synthesis of the geological evolution of New Zealand, it is possible to derive from the geology additions and refinements to the hypothesis. Thus quantitative time and space scales can be deduced for the various events associated with orogenesis. Time relationships during orogeny are of particular interest. In Fig. 3.13 I have detailed the time ranges of various events associated with the Rangitata Orogeny. Certain aspects of the incomplete record of the Permo-Triassic events show similar time ranges. These time relationships could also be compared with those deduced in other orogenic belts (see also Dewey, 1969b, fig. 3, for a generalised scheme) and, particularly if similar patterns are found elsewhere, used as initial guidelines in the investigation of lesser known areas. It is these predictive aspects of the plate-tectonic hypothesis which have led to perhaps the greatest revolution in the history of the earth sciences.

3.3: THE CENOZOIC EVOLUTION OF NEW ZEALAND

The development of the Australian/Pacific plate boundary in relation to the regional sea-floor spreading history has been discussed previously (Chapter 2.6). It was shown that about 900 km of relative lateral displacement has occurred across this boundary, which runs along the Macquarie Ridge and through New Zealand. All of this movement has occurred within the last 60 MY, and hence any synthesis of the Cenozoic evolution of New Zealand must be made in this context. The basis of the following synthesis is therefore (a) to establish the timing and amounts of movement involved, leading to palinspastic reconstructions of New Zealand, and then (b) to interpret the Cenozoic geology in the light of these reconstructions.

3.3.1: *DEVELOPMENT OF THE ALPINE FAULT*

Relative displacements between the Australian and Pacific plates have been determined from Figs. 2.25 to 2.30, and are shown in Table 3.1 both as 10 MY increments, and totals. [The bottom line is a projection of these movements to 10 MY hence]. Displacement is gradually increasing, and comparison with Fig. 2.33 indicates a general correlation between rate of relative motion across the plate boundary and migration of the Australian/Pacific rotation pole away from the boundary. Table 3 therefore establishes the basic movement chronology which constrains the development of this boundary through continental crust between the Macquarie Ridge and the region north of New Zealand (both previously discussed in Chapter 2.6). Most of the 900 km shift between the Lord Howe

TABLE 3.1: MEASURED DISPLACEMENT BETWEEN AUSTRALIAN AND ANTARCTIC PLATES, IN RELATION TO ALPINE FAULT

MY b.p.	AUS-PAC OFFSET			ALPINE FAULT			DIFFERENCE=DEFORMATION BY BENDING	
	increment (km)	total (km)	average rate (cm/yr)	increment (km)	total (km)	average rate (cm/yr)	increment (km)	total (km)
50	0	0	0	not active			0	0
40	50	50	0.5	"	"		50	50
30	150	200	1.5	"	"		150	200
20	300	500	3.0	100	100	1.0	200	400
10	400	900	4.0	350	450	3.5	50	450
0	500	1400	5.0	450	900	4.5	50	500
-10								

Rise and Campbell Plateau must be accommodated during the last 40 MY.

The Alpine Fault (Fig. 3.1) is one of the major structural features of New Zealand. In South Island it is clearly defined as a narrow fault-zone (Wellman and Willett, 1942; Suggate, 1963) across which Paleozoic and Mesozoic sequences are now dextrally offset by 480 km (Wellman, *in* Benson, 1952; Wellman, 1956). To the south it is a continuous tectonic feature with the Macquarie Ridge (Suggate, 1963, 1968; Hamilton and Evison, 1967; Christoffel and van der Linden, 1972), but in northern South Island splays into several faults (Wellman, 1955, 1956), some of which can be correlated across Cook Strait into North Island (Wellman, 1956; Lensen, 1958; Grindley, 1960; Suggate, 1963).

The Alpine Fault is still active, and displaces streams and post-glacial terraces (e.g. Wellman, 1953; Clark and Wellman, 1959). Hence it seems reasonable to conclude that this fault has taken up the most recent part of the 900 km displacement across New Zealand. Alternative schemes would imply at least a two-stage movement on the Alpine Fault, and it is considered unlikely that once established, strike-slip movement would cease and then begin again after a pause. Thus on the basis of Table 3.1, I have estimated that the Alpine Fault developed about 20 MY ago. This is in agreement with Wellman (1971), Christoffel (1971) and others who interpret the Alpine Fault as an Upper Tertiary feature.

Most of the remaining 400 km of displacement across New Zealand is therefore likely to have occurred prior to the Alpine Fault. In northern South Island, it could perhaps be argued that other faults could account for some of this displacement. However, these all merge

southwards into the Alpine Fault (see Fig. 3.1), and in southern South Island no other major faults are observed which could accommodate 400 km lateral displacement.

In reconstructing New Zealand as part of Gondwanaland prior to Cenozoic sea-floor spreading (Chapter 2.2.2) it was noted that about 900 km horizontal displacement was needed, and that as a consequence, elements of the Mesozoic New Zealand Geosyncline were essentially "straightened". It is postulated that the regional evolution took place essentially as a reverse of this operation, and that up to 400 km lateral displacement between the Lord Howe Rise and Campbell Plateau occurred prior to the Alpine Fault by internal bending and deformation through this zone of continental crust. As a consequence of increasing deformation (see Table 3.1) the Alpine Fault finally developed as a distinct break, and then took up most of the strike-slip displacement. This model is illustrated by Fig. 3.14 and Table 3.1, and its geological implications are examined in the following section.

3.3.2: *CENOZOIC GEOLOGICAL HISTORY*

The palinspastic reconstructions of New Zealand (Fig. 3.14) are the basis for discussion of the Cenozoic evolution. Five of these have been enlarged (50, 30, 20, 10, 0, MY) and an intermediate one at 5 MY constructed. Paleogeographic data is replotted on each as appropriate. I have used the paleogeographic maps and descriptions of Fleming (1962, especially figs. 6-10), with additional data from the 1:250 000 map series (New Zealand Geological Survey, 1959-68), for this purpose. Facies are generalised over the interval shown. It is also noted that the designation "land" may in some instances reflect absence of marine

sediments due to subsequent erosion, and consequently the indicated shorelines may only be approximate in places, particularly in the southern part of the maps. Each paleogeographic map is followed by a block diagram, which schematically illustrates the structural development in terms of the distortion of an originally rectangular block at 50 MY ago. The following discussion is largely a commentary on these pairs of diagrams (Figs. 3.15 to 3.26).

Erosion of the Rangitata Mountains continued from the Upper Cretaceous (Section 3.2.6) into the Early Tertiary (Dannevirke and Arnold Series, Fig. 3.15). The Pacific shoreline transgressed westwards during this time (compare Figs. 3.9, 3.15) and sediments pass upwards from sands and muds into shelf limestones with only minor clastics near the shores. On the peneplained land surface, coal measures continued to develop locally. Marine influence in the west also began to increase, with transgressions in Westland and Northland (Fig. 3.15). These followed the opening of the Tasman Sea which then presumably spread eastwards over the Lord Howe Rise and into New Zealand. Sea-floor spreading between Australia and Antarctica was only just beginning 50 MY ago (Fig. 2.25) and no deformations in New Zealand had yet occurred. This stage is shown as an undistorted block in Fig. 3.16.

By the Oligocene (Landon Series, Fig. 3.17) marine transgression reached a maximum. Limestones and fine silts dominate sedimentation, indicating that only small, low-lying land areas remained. Tectonic activity began in the Upper Landon Series, and the land area began to increase as uplift occurred and gentle folds developed (Fig. 3.18). By the end of this stage, folding and faulting was occurring in Nelson/

Southland, and the Alpine Fault was probably initiated in North Island (see below).

In the Lower Miocene (Pareora Series, Fig. 3.19) tectonic activity increased rapidly, marking the beginning of the Kaikoura Movement. Clastic sediments became more abundant, indicating emerging land and marine regressions, and volcanic activity commenced in North Island. The extent of deformation is shown in Fig. 3.20.

Significant bending of the block has occurred (corresponding to about 200 km offset, Table 3.1). The initial Alpine Fault line continues north from the Macquarie Ridge, and links up with the Vening-Meinesz Fracture Zone (see Figs. 2.27, 2.28). In North Island strike-slip movement probably commenced about 30 MY ago, associated with the "secondary spreading system" in the New Caledonia and Norfolk Basins (Fig. 2.35, Chapter 2.6.2). However, vectors about the Campbell Plateau/Lord Howe Rise rotation pole only became parallel to the fault-line about 20 MY ago (Figs. 2.27, 2.28).

Thus in South Island it is postulated that the earliest deformation was accommodated by "squeezing" across the future fault-zone, with development of tightening synclines and associated parallel *sinistral* faults (see Fig. 3.20 - the fault movement can be demonstrated by bending a pack of cards in a similar manner). Grindley (1958, Maps 1, 2) and Wood (1962) show several *sinistral* faults in Southland, which converge towards the present Alpine Fault, and intervening anticlines and synclines which tighten northwards and some become entirely pinched out by overthrusting. Two of these faults (Hollyford and Livingstone Faults) and the narrow Key Summit Syncline are shown in Figs. 3.19 and 3.20. Further south-east the Key Summit Syncline

opens out into the much broader Southland Syncline. Corresponding "squeezed" structures occur in Nelson/Marlborough (Beck, 1964), where the narrow Nelson Syncline is bounded by steep thrusts. All of these faults probably began as strike-slip faults, and during continuing squeezing many became high-angle thrusts as now observed.

In North Island volcanism began in the Lower Miocene. A chain of dominantly andesitic volcanoes erupted along the eastern flank of Northland and Auckland (Kear, 1959a; Thompson, 1960; Healy *et al.*, 1964; Schofield, 1967, 1968), and tuffs are common in associated sediments. In the regional context, these volcanics can be related to the west-dipping Benioff Zone behind the Kermadec Trench (see Fig. 2.28). In such settings, calc-alkaline rocks generally occur between about 200 km and 500 km behind the trench (Hatherton and Dickinson, 1969), with a transition from andesites through dacites to rhyolites away from the trench.

The "Northland Allochthon" (distinguished on Figs. 3.15, 3.17, 3.19 and 3.21) was also emplaced during the Miocene. This supposedly allochthonous unit forms a large part of Northland, and includes Upper Jurassic to Lower Cretaceous greywacke-type sediments and basic volcanics (e.g. Tangihua Volcanics) which are very similar to the East Cape sequence (e.g. Grindley *et al.*, 1959, pp. 47-9), which was previously interpreted to be associated with a Mesozoic trench (Chapter 3.2.3; 3.2.4). This unit occupies an anomalous paleogeographic position in the reconstruction (Fig. 3.2), as it occurs in the midst of typical shallow-water shelf sediments. Its contact with underlying rocks has always been regarded as obscure, and Temple (1972) suggested the whole sequence was derived from the east as a huge allochthon in the Late

Oligocene. Detailed inspection of stratigraphic relationships favours an Early Miocene age for final emplacement, though movement of the allochthon may have begun in the Upper Oligocene. At the base of the allochthon, the Onerahi Chaos-breccia rests on rocks as young as Lower Miocene (Kear, 1959b; Kear and Waterhouse, 1967). The allochthon and underlying rocks are both unconformably overlain by almost flat-lying Middle Miocene conglomerates and sandstones (Hay, 1969, map 3). The absence of Pareora Series sediments in the central East Coast region (see Fig. 3.19) may perhaps indicate the passage of the allochthon across this area, thus implying that it has moved at least 200 km westwards.

Gravity sliding due to isostatic uplift of the East Coast region related to the southern extension of the Kermadec Trench (Fig. 2.28) may have been the mechanism of emplacement. Stonely (1968) describes the detachment and emplacement of a series of smaller thrust-sheets in the north of the East Coast during the Upper Oligocene to Miocene, which may similarly be related to this regional uplift. The time-relations of formation of the allochthonous sequence (during an Upper Jurassic episode of subduction) and its emplacement (during a later, unrelated Miocene episode of subduction) may have wider implications for the evolution of "subduction zones" (e.g: Coleman, 1971) elsewhere.

In the Middle and Upper Miocene (Southland and Taranaki Series, Figs. 3.21, 3.22) the Kaikoura Movements continued to increase in intensity, and by 10 MY ago displacement totalled about 500 km (Table 3.1). Marine regression and increasing uplift of the land occurred, and sediments generally became coarser, even conglomeratic in places

(Fig. 3.21). The hypothetical block is now even more deformed (Fig. 3.22), and the synclinal structures and sinistral faults noted earlier would have reached their maximum development just prior to the breakthrough of the Alpine Fault from the north as a discrete fault, to link up with the Macquarie Ridge. Thus the Southland and Nelson structures become separated by an increasing dextral displacement. Further development of these structures may have occurred, due to "drag" and compression, so that many originally sinistral faults are now steep thrusts.

By this stage the spreading system in the Norfolk Basin had probably declined in importance (Chapter 2.6.2) and increasing subduction was occurring in the trench extending from Fiji to New Zealand (Fig. 2.29). The continuation of the Alpine Fault along the Vening-Meinesz Fracture Zone was slowly disrupted, and new faults progressively formed to link up with the Kermadec Trench. These faults developed initially as "splays" from the earlier Alpine Fault line. The Clarence Fault (Fig. 3.22) was possibly the first of these faults.

Volcanic activity in the "Auckland Arc" became more widespread during the Middle and Upper Miocene. Along the eastern flank, andesites gradually gave way to dacites, and by the early Pliocene (see Fig. 3.23) rhyolites are predominant (Kear, 1959a, fig. 3; Thompson, 1965; Schofield, 1968). This trend in volcanism is consistent with the suggestion that the Kermadec Trench was migrating eastwards, and as the Benioff Zone thus became *deeper* under North Island, volcanism changed from andesitic to dacitic to rhyolitic, similar to the trend observed *across* present arcs (e.g: Hatherton and Dickinson, 1969).

In the Pliocene (Wanganui Series, Figs. 3.23, 3.24) tectonic

movements continued to intensify even further. South Island was becoming mountainous, supplying coarse sediments. North Island was low-lying, with further marine transgression occurring, and volcanism continued. This stability may reflect the shift of activity from the original Alpine Fault to the splays to the east, more of which were developing (e.g.: Hope Fault, Fig. 3.24). The convergence of these faults in the Kaikoura region produced "fault-angle depressions" in which coarse conglomerates were deposited (Lensen, 1962). In North Island volcanism in the Auckland Arc was confined mainly to rhyolites along the eastern flank of the zone (Kear, 1959a), suggesting that the Kermadec Trench and Benioff Zone had shifted even further east (Fig. 3.29). The beginning of activity in the Taupo Volcanic Zone probably began about 5 MY ago (see below).

The Recent (Hawera) stage of development of New Zealand is shown in Figs. 3.25 and 3.26. Uplift in South Island has produced the Southern Alps, with peaks over 3000 m. The Alpine Fault forms a very prominent western edge to the mountains, and the differential vertical movement across the fault ranges up to 10-15 km (Mason, 1961). This uplift on the eastern side of the fault is an isostatic response to a small component of subduction of the Australian plate along the Fiordland margin (see Fig. 2.30 and Christoffel and Van der Linden, 1972, especially figs. 6, 7). In the north of South Island, more splays have developed or are incipient (Fig. 3.26) linking the Alpine Fault to the Hikurangi Trench. The Kaikoura region (Lensen, 1962) is being forced upwards as the continental crust of the Chatham Rise is folded back against Marlborough, and mountains up to 3000 m occur. It is interesting to contrast

the Fiordland and Marlborough regions. Both show the passage of the plate boundary from continental to oceanic crust, but in Fiordland it is a "clean" transition, whereas in Marlborough the relative motion across the plate boundary is dissipated across a number of faults above the end of a Benioff Zone, and no precise line can be drawn to represent the boundary.

In North Island the major Recent tectonic development is the Taupo Volcanic Zone. This is essentially a rift-structure which has filled with large volumes of volcanics (Grindley, 1960; Healy, 1962, 1964; Healy *et al.*, 1964). Rhyolites are most abundant, with minor occurrences of andesites, dacites and basalts, and extensive ignimbrites (Healy, 1962). Karig (1970a, b; 1971) has shown that the Taupo Rift is the southern extension of the Havre Trough, which has formed during the last 5 MY or so as an inter-arc basin, related to eastward migration of the Australian/Pacific plate boundary (see also Chapter 2.6.2). Thus the Taupo Volcanic Zone is an extension of the inter-arc basin into continental crust. Hatherton (1969) and Hatherton and Dickinson (1969) have shown a correlation between the compositions and distribution of Recent volcanics in North Island (including areas outside the Taupo Volcanic Zone) and depth to the Benioff Zone.

In Table 3.1 I have projected the history of the Australian/Pacific plate boundary to 10 MY hence, when a further 500 km of displacement is predicted. This is assuming that the Australian/Pacific rotation pole does not shift from its present position, where it has remained for about 30 MY (Fig. 2.33). In southern South Island the

Alpine Fault will continue, and further splays will develop. If the Kermadec Trench continues to migrate eastwards, the Taupo Volcanic Zone will extend further south. However, as the Chatham Rise continues to collide with the Marlborough coast, compression will occur in this area, preventing the Taupo rift from developing through South Island.

The effects of the violent earth movements that have occurred in New Zealand during the Tertiary are perhaps unique. They have resulted in the exposure of a wide range of rocks, so that virtually the whole geological record is seen in outcrop somewhere. The problem of unravelling these geological data into a cohesive story may seem difficult, but, beginning with a geometrical analysis of sea-floor spreading, it is possible to build up sequential reconstructions, and then to develop a tectonic synthesis. The synthesis both has a built-in feed-back, and is predictive. The synthesis presented here can therefore be tested for self-consistency, but, more importantly, can be independently assessed and if necessary modified or rejected, in the light of geological data yet to be collected.

CHAPTER 4

EVOLUTION OF THE CONTINENTAL MARGIN OF SOUTH-EAST AUSTRALIA

This chapter aims to relate the detailed geological history of the continental margin of south-east Australia to the regional tectonic history. The continental margin (Fig. 4.1) includes the ocean-facing Eucla, Otway and Gippsland Basins, and also the Bass Basin between Tasmania and Victoria. A considerable amount of geological information has been collected in these areas, largely as a result of petroleum exploration over many years. Much of this information is unpublished, but most of it is accessible, and it provides a good basis for a comprehensive tectonic synthesis of the region.

The general approach follows the lines discussed in Appendix A. (see especially Figs A.1, A.7). A review of the geology (Section 4.1) is followed by the development and evaluation of a tectonic synthesis, within the regional framework already established in Chapter 2 (Section 4.2).

4.1. GEOLOGICAL OUTLINE OF SOUTH-EAST AUSTRALIA.

As a prelude to any attempt at a synthesis of the regional evolution, it is necessary to determine as much as possible about the geological framework from the available data. This Section briefly describes and summarises the geology of south-east Australia. It is compiled from many sources, referred to specifically where

appropriate, but also providing much general background to the study.

The area described is part of the present continental margin of south-east Australia (Fig. 4.1). Parts of the marginal sedimentary basins lie onshore, and have been studied over many years, particularly with regard to coal and oil exploration. Results of this work are scattered throughout the literature. Important sources include the publications of the State Geological Surveys and Royal Societies of South Australia, Tasmania and Victoria, the Bureau of Mineral Resources, Geology and Geophysics, and the Journals of the Geological Society of Australia and the Australian Petroleum Exploration Association.

More recently exploration for oil has been directed particularly to offshore areas. Some of the results of this work are included in the above literature, but most are still unpublished. However, under the terms of the Petroleum Search Subsidy Acts (P.S.S.A. 1957-1969), much of the information is available on open-file at the Bureau of Mineral Resources, Geology and Geophysics (housed at Canberra, Australia). This information includes seismic, gravity and aeromagnetic surveys and well-logs from many companies. The largest contribution has been by the Esso-B.H.P. partnership, who found the first commercial offshore petroleum field in Australia in 1965, and have subsequently discovered and developed several other oil and gas fields. A summary of these operations is given by Richards and Hopkins (1969), and later results by James and Evans (1971) and Griffith and Hodgson (1971).

In quoting references in this Chapter, I have distinguished the material available under the Petroleum Search Subsidy Act from other

literature (which is referred to in the normal way). The former is referred to in the text by a geographic name for the survey or well, followed by the abbreviation SS = seismic survey, AMS = aeromagnetic survey, or WCR = well completion report. These references are compiled separately from the main list, under the three survey headings above, and include the name of company, and Bureau of Mineral Resources Petroleum Search Subsidy Section Library file number by which they can be located. Some of this data has been published as separate reports by the Bureau of Mineral Resources, and this is also identified by its "Publication Number" in the lists.

4.1.1 REGIONAL GEOLOGICAL FRAMEWORK

The major areas to be described are the Otway, Bass and Gippsland Basins (Fig. 4.1). Most of the geological information is from sub-surface, but these areas are adequately covered by seismic surveys, with control on the stratigraphy from well-logs. Less data is available from the Eucla Basin and the southern margins of Tasmania, and the geology of these areas will be referred to later during the synthesis as appropriate (Section 4.2).

The geological setting of the Bass Strait region is illustrated in Fig. 4.2. The nucleus of the Australian continent is formed of Precambrian and Paleozoic rocks, which have been previously discussed (Chapter 2.3). Following cratonisation of the Tasman Orogenic Zone of eastern Australia, it has remained a stable area until the break-up of Gondwanaland began in the Mesozoic (Chapter 2.4). During this time large sedimentary basins have developed on and around the continent. Four broad types of basin can be distinguished in Fig. 4.2:-

- (i) Carboniferous to Triassic cratonic basins (Sydney; Tasmania),
- (ii) Mesozoic cratonic basins (un-named; not shown beneath Tertiary of Murray Basin),
- (iii) Mesozoic marginal basins (Eucla; Duntroon; Otway; Bass; Gippsland),
- (iv) Tertiary marine basins (Murray; sediments covering most of (iii)).

These basins are either resting entirely on Precambrian to Paleozoic continental crust (i; ii; Murray) or build out onto oceanic crust (iii). Tertiary sediments (iv) cover large areas of both, indicative of wide-spread marine transgressions.

The geology of each of the three major Bass Strait basins (Otway, Bass, Gippsland) is now reviewed in turn. Much of the data is summarised as maps, cross-sections and stratigraphic columns (Figs. 4.3 to 4.11) and the following notes (Sections 4.1.2; 4.1.3; 4.1.4) complement these figures. Further information is also incorporated into the paleogeographic maps accompanying the synthesis. Igneous activity is also an important feature, but will be discussed separately later (Section 4.2.5). The object is to present a broad, general review of each basin, as a basis for the synthesis. Thus no attempt is made to cover every aspect in detail, or to specifically refer to all of the literature. The figures are compiled from almost all of the available data, and in the text reference is made particularly to papers which discuss the more important features. Some attempt has been made to indicate reliability on the figures, but they strongly reflect the amount and quality of work carried out in a particular area. For example, many more faults are evident offshore, but this is probably due to the

much more extensive and penetrative seismic surveys which have been carried out at sea.

4.1.2: THE OTWAY BASIN

The Otway Basin (Figs. 4.1; 4.2) lies within the States of South Australia and Victoria. It is an assymetric basin forming part of the continental margin, and is continuous with the Duntroon and Eucla Basins to the west, and the west Tasmanian coast to the south-east. It also connects with the Bass Strait to the north of King Island. A comprehensive account of the geology of the Otway Basin is the subject of a recent volume prepared jointly by the Geological Surveys of South Australia and Victoria (Wopfner and Douglas (eds.), 1971). More recent offshore data, largely from Esso, complements this volume.

The major structural features of the Otway Basin are shown in Figs. 4.3 and 4.4. Along the northern margin, Paleozoic basement directly underlies thin Tertiary sediments. The sediment wedge thickens rapidly seawards, until it is in excess of 6000 m beneath the edge of the continental shelf. Further offshore there is only poor seismic control on the basement, but the sediments must thin until only some hundreds of metres thick where they overlie oceanic crust in water depths in excess of 5000 m (c.f. Houtz and Markl, 1972, fig. 6). This transition from continent to ocean basin is also reflected by a rise in the depth of the Mohorovicic Discontinuity, from about 35-40 km beneath the Australian continent to about 10 km in the ocean. (e.g: Hawkins *et al.*, 1965).

The basement morphology is irregular, and several prominent "highs" and "troughs" are evident. These are reflected by the total sediment isopachs (Fig. 4.3), which onshore and in shallow water are roughly equivalent to

structure contours on basement. The principal features are the Robe and Penola Troughs in the west, and the Torquay Embayment in the east. The former are graben-like troughs which splay off the main trend of the margin and run onshore. They are bounded to the south by the Kalangadoo High, on which the lowest sedimentary units are absent. The Torquay Embayment trends north-easterly, roughly perpendicular to the margin. It is regarded as part of the Otway Basin only during the Lower Cretaceous, after which it is much more closely related to the Bass Basin.

The general basin stratigraphy is summarised in Fig. 4.5. Representative stratigraphic columns are shown from two areas of the basin. Elsewhere the succession is broadly similar, but local variations in nomenclature occur (c.f: Kenley, 1971, encl. 5-2). The stratigraphy is fully described in Wopfner and Douglas (1971).

Economic basement (defined as non-prospective for petroleum) comprises the Paleozoic metamorphics and granites of the southern end of the Tasman Orogenic Zone, which outcrop along the northern margin of the basin. Elsewhere basement has been intersected in wells near the margin, and in a few deep wells nearer the centre (e.g: Fergusons Hill, Kalangadoo, Robertson -1 WCR's). The data are insufficient to determine the southern limit of the Paleozoic basement, beyond which sediments presumably rest on oceanic crust.

The oldest basin sediments recognised are Jurassic, lying below the more extensive Otway Group (c.f: Wopfner *et al.*, 1971, pp. 391-2). They are known only from subsurface intersections, and are only poorly inter-correlated. In the Penola Trough (Fig. 4.3) the informally named "Casterton Beds" are the lowest unit (c.f: Casterton, Penola, Robertson-1

and -2 WCR's), comprising at least 400 m of sandstone, shales, minor coals and basalts deposited in a freshwater environment. Similar rocks also occur at the base of the succession in Victoria (c.f: Hawkesdale, Woolsthorpe WCR's), but were probably not continuous across the basin.

The next unit recognised is the Crayfish or Pretty Hill Sandstone, which is generally a fairly mature, porous quartz sandstone. (c.f: Crayfish, Garvoc, Pretty Hill, Woolsthorpe WCR's). In the Penola Trough this unit is not distinguished, but its equivalent is probably part of the Otway Group (Lower Unit). The apparent variation in pre-Otway Group sediments may in part be due to difficulties in correlation, but probably also reflects the presence of several distinct depositional areas during the early stages of basin evolution.

The Otway Group (c.f: Rochow, 1971b) is a much more widespread unit, probably up to 3000 m thick in places. Outcrops form the highlands of the Merino Uplift and Otway Ranges. It is a Lower Cretaceous, non-marine sequence of rather monotonous arkoses, greywackes, silts, mudstones and some coals, derived from rapid erosion of highlands to the north, and also probably the south (Otway ER-68 SS, p. 4).

The Upper Cretaceous Sherbrook Group is divided into several formations (c.f: Glenie, 1971). The sequence passes up from a basal transitional sandstone (Waare Formation) into marine mudstones (Belfast Formation) and eventually into prograding sandy beds which continue into the Tertiary. The total Upper Cretaceous section is up to 4000 m thick in the central part of the basin. Detailed studies (e.g: Taylor, 1971) indicate a progressive marine transgression from the west during

the Upper Cretaceous. Taylor (1964) and Bock and Glenie (1965) recognise a series of several transgressive-regressive cycles in the Upper Cretaceous and Tertiary of the present onshore areas. The Tertiary section reflects these cycles, and comprises a variety of sediments, passing upwards generally from sands to marls and limestones. An important regional unconformity, diachronous from Paleocene to Eocene, is also recognised (Richards and Hopkins, 1969).

The structural and stratigraphic framework of the Otway Basin is further illustrated in Fig. 4.4, which is a series of composite profiles roughly normal to the trend of the margin. Several features are evident from these profiles (see also Rochow, 1971a, pp. 313-5):-

- (a) Thick pre-Otway Group sediments occur only in the Robe and Penola Troughs, and in the western offshore part of the basin (profiles 1-5).
- (b) The Kalangadoo High forms a prominent basement ridge during pre-Otway Group times, but the Otway Group is deposited in a south-thickening wedge right across it (profiles 1-6). Otway Group sediments are extensively down-faulted to the south (profile 1).
- (c) Upper Cretaceous sediments also thicken generally southwards, and are cut by syn- and post-depositional faults. On several seismic lines (not shown) the sediments clearly occur in down-to-the-south fault depressions, dipping back towards the north. The Upper Cretaceous section is thickest in the centre of the basin (profiles 4-10), elsewhere it is much thinner.

- (d) Tertiary sediments form prograding wedges, up to about 2000 m thick (profiles 1-12). The thickest part (profiles 9-11) is further east than the Upper Cretaceous maximum. Very little fault movement is evident in the Tertiary, particularly above the Paleocene to Eocene unconformity.
- (e) The Torquay Embayment has a thick section of Otway Group sediments, overlain by thin Upper Cretaceous and Tertiary (profiles 13-14).
- (f) The south-eastern part of the basin, off King Island (profiles 15-16), and its continuation along the western margin of Tasmania, is much narrower than the main part of the basin in South Australia and Victoria.

4.1.3: *THE BASS BASIN*

The Bass Basin (Figs. 4.1, 4.2) lies between Victoria and Tasmania. It is entirely surrounded by outcropping or shallowly submerged Paleozoic rocks, except in the north-west where it connects with the Torquay Embayment and the Otway Basin. Apart from thin onlapping Tertiary sediments in northern Tasmania, it is also wholly submerged. All of the work in the offshore basin has been carried out by Esso-B.H.P., and is summarised by Richards and Hopkins (1969).

The structural framework of the Bass Basin is shown in Figs. 4.6, 4.7. It is a roughly elliptical, NW-SE trending basin, with up to 6000 m of sediments in the centre and south-east. The north-eastern margin is formed by the Bassian Rise, a ridge of Paleozoic granite extending from Tasmania to Wilson's Promontory, which breaks the surface as several small

islands. It is probable that this ridge continues through to the major Paleozoic outcrops in Victoria, with only a thin cover of Mesozoic in the south-western part of the South Gippsland Highlands, and hence that the Bass and Gippsland Basins have never structurally connected (see Section 4.1.4 for more detailed discussion). On the western flank of the basin, shallow basement extends beneath Tertiary sediments some distance to the north of King Island. During the Lower Cretaceous the Torquay Embayment received sediments very similar to those of the Otway Group to the west (see Fig. 4.5, profiles 13, 14). However, in the Upper Cretaceous this connection was broken, and the sediments in the Embayment are much more like those of the Bass Basin. A shallow ridge then probably extended from King Island to the Otway Ranges in southern Victoria, acting as a barrier to the entry of Upper Cretaceous seas from the Otway Basin.

The stratigraphy of the Bass Basin (Fig. 4.8) is based on only three subsidised well-logs (Bass-1, -2 and -3 WCR's) and data from the Torquay Embayment (c.f: White, 1968). Economic basement around the basin margin varies from Precambrian (on King Island) to Triassic (in Tasmania), and it is likely that similar rocks underlie the centre of the basin. The two wells which reached basement (Bass-2, -3 WCR's) bottomed in altered mudstone (?Mesozoic) and metamorphic rock respectively.

The oldest sediments drilled are Upper Cretaceous, but it is thought that Lower Cretaceous sediments occur in the deeper parts of the basin, particularly in the south-east (Richards and Hopkins, 1969, p. 4; Bass B69A SS, p. 3). Up to 3000 m of Upper Cretaceous to Eocene sediments, informally named the "Eastern View Complex" by analogy with

similar aged sediments in the Torquay Embayment (the Eastern View Coal Measures, c.f: White, 1968, pp. 84-5) are a deltaic-like complex of interbedded sandstones, siltstones, shales and coals, derived largely from the south and east sides of the basin (Richards and Hopkins, 1969, p.5). An unconformity developed across the basin in the Upper Eocene, and was followed by a marine transgression from the west. The rest of the Tertiary section consists of shales, sandstones, minor tuffs and calcareous sediments, altogether up to 2000 m thick. Sediments were not deposited on the Bassian Rise until the Miocene (Bass B69A SS).

The basin structure is illustrated by a series of profiles in Fig. 4.7. The deepest part of the basin is formed by large faults downthrowing the basement (profiles 9-11), forming grabens filled with thick pre-Oligocene sediments. In the centre part, basement reflections are not identified, and at least 5000 m of sediments occur (profiles 4-8). The profiles also show that very little fault movement post-dates the Eocene unconformity.

Beneath the centre of the Bass Basin the crust is thinner than normal continental crust. Interpretation of the results of the Bass Strait Upper Mantle Project seismic refraction experiment (Johnson, 1972, chapter 5) indicates that the crust under Bass Strait is only 20-25 km thick, compared to 30-35 km under Tasmania and up to 45 km under Victoria. A similar conclusion was deduced from a marine gravity traverse conducted by the Bureau of Mineral Resources (P. Cameron, pers. comm.). These results tentatively suggest the crust is thinnest beneath the deeper parts of the basin, probably as a result of crustal stretching.

4.1.4: THE GIPPSLAND BASIN

The Gippsland Basin (Figs. 4.1, 4.2) lies in the eastern Bass Strait between Victoria and Tasmania, and extends onshore in eastern Victoria. The onshore part of the basin includes the Latrobe Valley Coalfields, where huge reserves of brown coal are exploited for electricity generation in Victoria (c.f: Gloe, 1960). Near the coast, oil exploration activity has been conducted for many years, and small uncommercial oil and gas fields occur at Golden Beach and Lakes Entrance. Offshore exploration has been conducted largely by Esso-B.H.P. Australia's first offshore well (Barracouta-1) was drilled in 1964/5, and later developed as a commercial gas producer. Subsequently Esso-B.H.P. have discovered and developed several other oil and gas fields in the basin, which are now supplying a substantial proportion of Australia's requirements (c.f: Griffith and Hodgson, 1971; Stratton, 1973).

The structural framework of the Gippsland Basin is outlined in Figs. 4.9, 4.10. The basin is roughly triangular in shape, opening eastwards towards the Tasman Sea. Paleozoic rocks of the Tasman Orogenic Zone outcrop on both the northern and southern flanks. The Bassian Rise on the southern side appears to have completely separated the Gippsland and Bass Basins until the Upper Tertiary, and only a narrow connection has existed with the Otway Basin (see below). The Tasman Sea floor to the east is normal oceanic crust (Officer, 1955), with only a few hundred metres of sediments (P. Cameron, pers. comm.). The basin sediments thicken from the flanks to over 6000 m in the basin centre, and thin eastwards onto the Tasman Sea floor.

The stratigraphy of the Gippsland Basin is summarised in Fig. 4.11.

The succession differs from that in the Otway and Bass Basins in that the upper formations are quite strongly diachronous. The three columns (of Fig. 4.11) together thus illustrate a "time cross-section" of the Gippsland Basin (see Hocking, 1972, fig. 3), showing the succession in the Latrobe Valley Coalfields, the coastal area of Gippsland, and offshore, respectively.

Economic basement is reached in only a few wells, and is similar to the Paleozoic rocks which outcrop around the margins. Two wells on the northern side (South-west Bairnsdale, Duck Bay WCR's) encountered a few hundred metres of Devonian and Permian sediments (respectively) lying above true economic basement beneath the Mesozoic succession. These sediments are probably relicts of Paleozoic platform cover.

The oldest basin-filling unit is the Jurassic(?) to Lower Cretaceous Strzelecki Group. On the northern edge of the basin, in the Latrobe Valley, the "Tyers Group" is a basal unit of up to 600 m of conglomerates, greywackes and sandstones, derived from rapid erosion, and deposition in alluvial fans, of Silurian and Devonian rocks to the north (Philip, 1958). The major part of the Strzelecki Group consists of arkoses and mudstones, with minor grits, conglomerates and coals (c.f: Edwards and Baker, 1943). The thickest drilled section is 2400 m (Wellington Park WCR), but over 3000 m may be present in the centre of the basin. The Strzelecki Group is lithologically similar to the Otway Group (c.f: Edwards and Baker, 1943), being derived largely from erosion of Paleozoic sedimentary, metamorphic and igneous rocks.

This similarity has led to a commonly expressed view that a major east-west trough developed right across south-east Australia in

the Jurassic to Lower Cretaceous. However, this view is rather too simple - the similarity in lithology may reflect a similar provenance for the sediments, but does not provide any evidence for a single basin. In the Gippsland Basin, the Strzelecki thins westwards, and is exposed in the South Gippsland Highlands (Fig. 4.9). Several Paleozoic outcrops occur along the south-western side of these hills, from the Mornington Peninsular through to Wilsons Promontory, with gaps of no greater than 35 km between outcrops. The Mornington Peninsular is a ridge extending south from the main Paleozoic outcrops of Victoria (Keble, 1968). Wilsons Promontory is the northern end of the Bassian Rise, which extends to Tasmania. It is unlikely that uplift (and consequent stripping off of younger sediments) of more than a few hundred metres has occurred in South Gippsland, as the younger sediments wedge out towards the Highlands (c.f: Hocking, 1972, figs. 1, 4), and the major tectonic regime appears to have been dominantly tensional (see Section 4.2). Hence any east-west connection must be in the narrow gap south-west of the South Gippsland Highlands. At least 1000 m of sediments are recorded on the coast on this region (Tarwin Meadows No. 1 well, P. Bollen, pers. comm.) suggesting that the basin continues in a south-westerly direction offshore. However, aeromagnetic interpretation (Bass Strait and Encounter Bay AMS) indicates that the basement rises to less than 500 m further offshore, before it deepens again west of the Mornington/King Island High into the Torquay Embayment, and southwards into the Bass Basin. Hence it appears that the Gippsland Basin was separated by shallow basement ridges from both the Otway and Bass Basins throughout its evolution.

The Latrobe Group (James and Evans, 1971) unconformably overlies

the Strzelecki Group and basement in the offshore part of the basin, and ranges from Upper Cretaceous to Eocene. It is a diachronous unit, and onshore its equivalents are the Paleocene to Eocene Latrobe Group, and Eocene to Miocene Latrobe Valley Coal Measures (c.f: Hocking, 1972). The whole Latrobe Group is a major deltaic complex, derived largely from the north and west, and consists of non-marine assemblage of sandstones, coals, mudstones, siltstones and shales, typifying the varied nature of deltaic sediments (Richards and Hopkins, 1969; Franklin and Clifton, 1971; James and Evans, 1971, Bein *et al.*, 1973). It is up to 4000 m thick in the centre of the basin, where it overlies the Strzelecki Group, but spreads out over a larger area and onlaps basement (Figs. 4.9, 4.10).

Within the Latrobe Group in the eastern part of the basin, two large channel systems (the Marlin and Tuna/Flounder channels) have been mapped (Richards and Hopkins, 1969; James and Evans, 1971). These channels were cut into the top of the Latrobe Group during the Eocene. They were subsequently filled by marine incursions from the east, distinguished separately as the Flounder and Turrum Formations (James and Evans, 1971). Hocking (1972) considers these formations to be part of the Seaspray Group, in which he includes all the marine sediments in the eastern Gippsland Basin. The Flounder and Turrum Formations are both composed of up to about 500 m of shales and some clastics, probably representing the "mixing" of channel-filling detritus and incoming marine sediments. At the top of the Latrobe Group, James and Evans (1972) also recognise up to 40 m of shallow marine muds and silts, comprising the Gurnard Formation.

A marine transgression from the east occurred during the Oligocene and Miocene. The first major unit is the Lakes Entrance Formation, which consists of up to 500 m of mudstones lying unconformably on the Latrobe Group (offshore), and 250 m of sandstones, muds and gravels conformable with the Latrobe Group further west (onshore). This unit is followed by the Gippsland Limestone, comprising up to 1500 m of limestones and marls, with clastics and mudstones filling incised submarine canyons (James and Evans, 1971). The final phase of sedimentation is a return to clastic deposition (Sale Group) in the onshore parts of the basin (Hocking, 1972).

The structure of the basin is illustrated by a series of cross-sections (Fig. 4.10). These show clearly that the Strzelecki Group occupies a much smaller part of the basin than the succeeding units. Hocking (1972) distinguishes a "Strzelecki Basin" from the overlying "Gippsland Basin", using the latter term only for the Upper Cretaceous and Tertiary sediments. James and Evans (1971) recognise "three major areas of the (Gippsland) basin, separated by bounding fault complexes - these are a central deep basin (i.e: the Strzelecki Basin), and northern and southern "platforms". The "Strzelecki Basin" trends east-west, and is a half-graben formed by down-to-the-north faults along its southern margin (Fig. 4.10, profiles 4-6; Hocking, 1972). Sediments thin out both onto the northern margin, and west towards the South Gippsland Highlands. Its development was contemporaneous with the early Otway Basin, but the two were most probably distinct, separate basins as already discussed. Further east (profiles 7-10) there is only poor data available on the thickness of the Strzelecki Group, but it appears to thin out again and narrow (c.f: Richards and Hopkins, 1969, fig. 4). Hence it is tentatively suggested that the "Strzelecki Basin" was

initially an elongate depression, closed on at least three sides by shallow basement.

The overlying Latrobe Group delta developed initially within the limits of the earlier "Strzelecki Basin". The depocentres of various units of the Group moved gradually westwards during the Upper Cretaceous to early Eocene (Richards and Hopkins, 1969; Hocking, 1972). The southern margin was still fault-controlled during this period (profiles 4-8). In the Upper Eocene, fault-movement diminished, and the delta transgressed onto the southern basement platform. By the Miocene, true (non-marine) deltaic facies are confined to the Latrobe Valley region in Victoria. Channel-cutting began in the Eocene in the eastern part of the basin. The present top of the Latrobe Group is partly a topographic (erosional) surface resulting from this activity. The marine Lakes Entrance Formation and succeeding units transgressed westwards across this surface, and onto the basin flanks.

Major faulting in the basin appears to have ceased during the Eocene, prior to deposition of the Lakes Entrance Formation. Subsequent movements seem to have been relatively small, and confined within the limits of the initial "Strzelecki Basin" (Hocking, 1972). Further movements in the area have resulted in "draping" of the Lakes Entrance Formation and succeeding units over older blocks, and the formation of broad anticlinal structures. Some of these contain oil and gas, trapped in porous Latrobe Group sandstones beneath a cap-rock of Lakes Entrance Formation mudstones (c.f: Griffith and Hodgson, 1971).

Richards and Hopkins (1969, p. 4) suggest that a large wrench-fault system and anticlinal folds developed in the northern half of the

basin in the Upper Tertiary, possibly "associated with a right-lateral shear acting in the northern portion of the basin". However, there is little evidence for such implied major movements. The concept of such a major shear was put forward by Carey (1968, p. 5) who suggested that a 100 km dextral displacement had occurred between Tasmania and Victoria. If this were an Upper Tertiary feature, then some evidence of it would be expected in the Otway Basin to the west, where however virtually no tectonic activity has occurred since the Eocene (Section 4.1.2). Thus the concept of "draping" over *older* fault-controlled blocks to produce the anticlinal structures, rather than compressional folding, seems a more reasonable hypothesis. Hocking and Taylor (1964) and Hocking (1970) give evidence to support this hypothesis from the onshore part of the basin. This point is discussed again later.

4.2: TECTONIC SYNTHESIS OF SOUTH-EAST AUSTRALIA - DEVELOPMENT AND EVALUATION

The preceding review of the geology of the three Bass Strait basins provides the basis for development of a regional synthesis. The broad tectonic setting has already been examined in Chapter 2, and the chronology of major events established. The synthesis of south-east Australia involves *integration* of the regional and local aspects, and the development of a model which is equally compatible with both. This approach also allows for further evaluation of the conclusions of Chapter 2 against specific geological data. If the previously constructed model is at variance with the local data, then it may be possible to modify it, until an acceptable synthesis is made. The general principles underlying this approach are more fully discussed in Appendix A.

The development of the synthesis begins with a review of the present regional framework and the pre-drift reconstruction of Australia and Antarctica (Section 4.2.1). A structural analysis of the region (Section 4.2.2.) is followed by an outline of the geological evolution in chronological sequence (Sections 4.2.3; 4.2.4; 4.2.5). Igneous activity is an important aspect of the basin evolution, but this part of the discussion is more conveniently deferred to a separate section (4.2.6). Finally an attempt is made to evaluate the synthesis and to discuss some predictions and conclusions arising from it (Section 4.2.7).

4.2.1: REGIONAL TECTONIC FRAMEWORK AND THE AUSTRALIA/ANTARCTICA RECONSTRUCTION

The present tectonic setting of south-east Australia is shown in Fig. 4.12. The region is bounded to the south and east by oceanic crust, and lies over 900 km from any active plate margin at present. Magnetic anomalies indicate that the Tasman Sea floor is about 80 MY old near to the eastern coast of Australia, and the south-east Indian Ocean floor about 55 MY old just to the south of Australia. The mid-ocean ridge between Australia and Antarctica is the site of present sea-floor spreading. South of Tasmania, this ridge is offset by a series of transform faults. These displace the ridge axis progressively southwards, such that its general trend parallels the southerly bend in the Australian and Antarctic margins. These faults extend towards the continental margins as aseismic fracture zones, but these lose their identity near to the base of the continental slopes.

The south-east Australian continental margin is shown in more detail in Fig. 4.13. The morphology of the region has been described by several authors (Von der Borch, 1967; Conolly, 1968; Conolly *et al.*, 1970; Von der Borch *et al.*, 1970). The Tasman Sea margin is narrow and relatively steep.

Large submarine canyons occur in the eastern Bass Strait, incised into the sediments of the Gippsland Basin (Conolly, 1968). The East Tasman Plateau and South Tasman Rise are probably small fragments of continental crust partly rifted away from Tasmania (Hayes and Conolly, 1972, pp. 137-8). The southern margin of Australia is generally broader, and the slope more general. Two prominent "marginal plateaus" (the Ceduna and Beachport Plateaus) occur, and also several submarine canyons (Von der Borch, 1967; Conolly, *et al.*, 1970; Von der Borch *et al.*, 1970).

The mapped marine magnetic anomaly pattern (Weissel and Hayes, 1972, fig. 2) is also shown. In the Great Australian Bight, anomalies up to 22 (54 MY) are observed close to the base of the continental slope. Further east, to the north of where the ridge is offset by a succession of closely spaced transform faults (Fig. 4.12) the magnetic pattern is more obscure near the margin. The boundary of the "magnetic quiet zone" (c.f: Weissel and Hayes, 1972) roughly corresponds to the 5000 m bathymetric contour, suggesting perhaps that sediments on the lower continental slope may be blanketing any anomalies present north of this line. Very closely-spaced profiling will probably be necessary to further delineate the magnetic pattern in this area.

South-east Australia is also a region of minor seismic activity, and recorded earthquakes greater than Magnitude 3 are shown on Fig. 4.13. Some of these are associated with specific faults, whilst others are apparently more random. This pattern of activity probably results from continuing epeirogenic readjustments within the Australian plate, rather than delineating incipient plate boundaries as suggested by Cleary and Simpson (1971), as there is no supporting evidence (e.g: developing rift valleys) to suggest that the Australian plate is about to break up.

The pre-drift reconstruction of Australia, Antarctica and continental fragments to the east has already been discussed (Chapter 2.2; Fig. 2.8). This reconstruction was assembled on a scale of 1:15 million, using the 2000 m bathymetric contour as a base, and it was shown that major geological elements were thus brought into alignment (Fig. 2.9). In this Chapter, the reconstruction of Australia and Antarctica has been re-examined with particular regard to the detailed geology of the opposing coasts, and an independent "fit" obtained (Fig. 4.14) on a 1:5 million scale. This differs from those of Sproll and Dietz (1969) and Smith and Hallam (1970), which were based on morphological matching of selected isobaths using a computer, in that Australia is shifted about 250 km to the left (anticlockwise) with respect to Antarctica. This fit is considered more realistic for the reasons outlined below. However, it is pointed out that the total length of "contact" between Australia and Antarctica is about 3500 km, and the difference of 250 km will therefore have little apparent effect on large-scale Gondwanaland reconstructions. [Note: the fit preferred here is used in the final version of Fig. 2.8]. But, on a local scale, the difference has important consequences and implications which previously seem to have been largely neglected in morphological fits between continents, whether constructed by computer or by eye.

Morphological fits are based on the selection of a present bathymetric contour. There is thus an implicit assumption that the selected contour is a reflection of the *original* shape of the continental edge - that is, the width of later accumulations of sediments is roughly constant along the margin. This assumption is not necessarily valid though, as

wide sediment wedges can build out in localised spots, such as near major deltas. Sproll and Dietz (1969) point out that there is a substantial overlap of the Antarctic 1000 fathom (approx. 2000 m) isobath onto the Australian margin in the vicinity of the western Otway Basin. In explanation, they suggest (op.cit.) that either the Antarctic bathymetry is in error, or that other causes have modified the original shape of the continental margin on one or both sides.

There are few data available for the Antarctic margin in this area, but it seems unlikely that errors in the bathymetry of the magnitude required could occur. The effect of the ice-load on the Antarctic continent may also be important, as if the continent were depressed, the "equivalent" contour to 2000 m on the Australian side might be deeper, say 3000 m or more. However, in this area the ice-cover is thin (Thiel, 1962, fig. 5) and the crust of normal continental thickness (Woollard, 1962, fig. 10), so that any ice-load effect is probably insignificant. In any case, the 3000 m and 2000 m bathymetric contours are close together along this part of the margin. Antarctica is surrounded by a land-derived wedge of sediments, possibly up to 4000 m thick (Houtz, *et al.*, 1971), but there is as yet no evidence to suggest that the entire "bulge" is composed of such sediments. Hence it is tentatively assumed that part at least of this bathymetric feature is older continental crust. In contrast to the Australian margin, it is noted that basement is exposed right up to the coast of Antarctica, with no evidence of younger cover remaining onshore.

The Australian margin has already been discussed in detail (Section 4.1). In the Otway Basin (see Figs. 4.3, 4.4) the thick part of the sedimentary accumulation is much broader than to the west, or off the

western coast of Tasmania, and hence the bathymetry is not a true reflection of the original shape of the margin. It is suggested therefore that definition of the continental edge might be more accurately made by using an *isopach* within the continental margin sediments, rather than present bathymetry. Thus in Fig. 4.14 I have delineated the 3000 m isopach of Mesozoic/Cenozoic sediments along the Australian margin, and used this line as the basis for morphological fit with Antarctica. The result is a fairly good fit, displaced some 250 km from that of Sproll and Dietz (1969), which largely eliminates the 'problem of the Antarctic "bulge" which now fits into the Otway Basin.

This postulated fit is rather tentative, and points to the need for more data concerning the bathymetry and sediment thickness along the Antarctic margin. It does however suggest one possible close geological correlation between Australia and Antarctica which could be investigated. The Dundas Trough of Tasmania has been interpreted as a Cambrian subduction zone (Solomon and Griffiths, 1972). It runs off the western side of Tasmania, and hence it might be expected to continue in Antarctica. The fit of Sproll and Dietz (1969) would place its likely extension to the east of Victoria Land, submerged in the Ross Sea. The fit suggested here takes its likely extension onto the Antarctic coast.

The Dundas Trough is characterised by a belt of volcanics and ophiolites in the Cambrian, followed by a post-collision sequence of conglomerates and later shelf-sediments in the Ordovician (Solomon and Griffiths, 1972). The Trough lies between Precambrian blocks. In Victoria Land, there is a strip of Lower Paleozoic rocks (the Bowers Group) exposed between flanking Precambrian masses (Gair *et al.*, 1970). The

sequence consists of Cambrian to Ordovician greywackes, volcanic-derived argillite, volcanic agglomerates and quartzite conglomerates (Sturm and Carryer, 1970; Laird, pers. comm.). This sequence and its setting is grossly similar to that of the Dundas Trough, and it may thus be the southern continuation of the latter. This suggested correlation is highly speculative at this stage, but, if correct, evidence of ophiolites and the volcanic arc might be expected beneath the exposed Bowers Group sediments. This prediction could be tested either by geophysical surveys (e.g: aeromagnetic, gravity) or by drilling.

The oldest oceanic crust adjacent to south-east Australia is about 80 MY (Upper Cretaceous), but fragmentation of the Gondwanaland continent in the region had begun much earlier. The reconstruction (Figs. 2.8, 4.14) provides the basis for the study of the early evolution, in as much as during this stage, the continents remained fairly closely held together, and events recorded in the geological history of either side should be inter-related. Geological data are only available from the Australian margin, and hence the region synthesis is constructed around this information. It does however lead to a predicted history of the Antarctic and Lord Howe Rise margins, which in the future could be studied in detail, and thus allow an evaluation of the present synthesis.

4.2.2: *REGIONAL STRUCTURAL ANALYSIS*

Rifts are generally considered to be the result of crustal tension, which in the brittle upper layers is relieved by normal faulting and consequent downthrow of blocks between faults. In the analysis of a region characterised by rifting, the first problem is perhaps to determine the regional stress-pattern which has created the rifts, and to ascertain

whether all the observed structures can be related to a simple overall direction of tension. In terms of the sea-floor spreading hypothesis, it is supposed that rifts are the early stages in continental break-up (e.g: Carey, 1958; Dewey and Bird, 1970b; Griffiths, 1971b), and hence that the rift pattern should be related to the direction of pull-apart between the two (or more) plates derived from the fragmentation of a larger plate. If such a simple stress-system operates, then two further points must be considered:-

- (i) Does the regional principal stress orientation remain constant, or vary?
- (ii) Do older basement structures control the local stress distribution?

In areas where continents are now separated by ocean basins, the rift-forming stress systems can only be studied by analysis of the structures which they have produced. Thus in south-east Australia, now a long way from any active plate-margin, the original stress patterns must be deduced from study of the faults which have controlled the basin evolution. It is necessary to first derive local stress patterns, and then to determine whether these can be interpreted in terms of a single regional tensional direction, perhaps modified by local controls and varying in time.

As a first step, the orientation of all faults in the Otway, Bass and Gippsland Basins (see Figs. 4.3; 4.6; 4.9) has been measured, and the results plotted as rose-diagrams (Fig. 4.15). No attempt has been made to determine the timing of movement on individual faults, but it was previously shown (Section 4.1) that there has been little significant fault-movement in any of the basins since the Eocene. Thus the analyses represent the sum of all the observed faults active between the Jurassic and early Tertiary.

Neither has any attempt been made to statistically sample the fault trends. The number of faults mapped is very largely dependent on the spacing and penetration of seismic surveys, and thus the distribution of measured fault trends is very uneven, making systematic sampling impossible. Hence the rose diagrams (constructed as indicated in Fig. 4.15) have no quantitative significance, and only emphasise the general trends of basin-forming faults.

The rose-diagrams for each basin (Fig. 4.15 upper) show a dominant fault-orientation in one sector (and its reflection). Thus in the Otway Basin the trend is between N and W; in Bass between N and NW; and in Gippsland between NW and SW. Each basin also shows a pronounced minimum roughly normal to these trends. Summation of all the faults (Fig. 4.15, lower), although biased by the Otway trends, shows that virtually all the faults in the region are restricted to the quadrant between N and W, with a pronounced minimum in a NNE direction.

Before attempting to interpret the rose-diagrams in terms of stress-directions, it is necessary to consider the type(s) of faults involved. Anderson (1951, pp. 13-16) distinguishes three fundamental types of faults, based on the orientation of three mutually perpendicular stress-directions. These are classed as normal, thrust and wrench faults (see Fig. 4.16). The type of fault developed is related to the orientation with respect to the earth's surface of the maximum, intermediate and minimum principal stresses. These stresses are conventionally regarded as *compressional*, and so the minimum compressive stress is equivalent to the maximum *tensional* stress. In dealing with rift regimes, this stress (T_3 in Fig. 4.16) must be one of the horizontal stresses, as the vertical principal stress is always compressive (due to gravity). Hence only normal and wrench faults are likely in such a regime. The strike of normal faults (Fig. 4.16) is always perpendicular

to T_3 , and hence for all purely normal faults, the local orientation of T_3 can be determined. Strike-slip (wrench) faults form conjugate pairs inclined at about 45° to T_3 - i.e: T_3 bisects such pairs (although only one set may be developed in an area).

Inspection of basin-forming faults in south-east Australia shows that the majority have a normal component, downthrowing basement blocks towards the basin centre. Most of these faults are mapped from seismic profiling, and it is impossible to readily determine if a strike-slip component is present. But since the maximum total extension across any basin is only about ten to twenty kilometers, it seems unlikely that major strike-slip faulting has occurred, especially on any single fault. Hence it is concluded that the rose-diagrams (Fig. 4.15) show essentially the trends of normal faults, and so the principal tensional stress (T_3) can be determined from these diagrams.

The "average" direction of T_3 can be deduced for each basin, and for the summation. T_3 for each individual basin lies in the N-E quadrant and the regional average is about $N35^\circ E$ (see Fig. 4.15). It can therefore be tentatively assumed that this latter direction represents the overall maximum tensional stress acting across south-east Australia during the Jurassic to Early Eocene rifting stage. It is shown below that this stress direction is compatible with more detailed analysis of the basin-forming faults on a local scale, and thus that the assumption is probably justified. It is also noted that any strike-slip faults, inclined at about 45° to this stress, would plot inside the envelope of fault trends in the rose-diagrams.

The most important point noted here is that the derived principal

tensional stress direction is about 45° difference from that of the fracture zones in the ocean south of Tasmania, which trend about $N10^\circ W$ (see Fig. 2.4) and record the direction of sea-floor spreading. This difference is considered significant, and suggests a non-relationship between the stress field which forms the rift valleys, and the direction of subsequent sea-floor spreading in those rifts which develop into ocean basins.

The total crustal extension across a basin can in theory be determined if both the basement and dip and throw of faults are accurately mapped right across the basin, and hence the faults which displace the basement could be "reversed", and the original length of the section determined. This is rarely possible in practice, as often the basement is not reached by seismic profiling. A different method can be used to estimate roughly the crustal extension in such cases. This is based on an approximate determination of the volume of sediment filling a basin, and calculation of the "shortening" required to close up the sediment-filled hole in the basement. It is assumed that the mantle rises to compensate for the basement downwarp due to faulting near to the surface. It is also assumed that isostatic equilibrium is maintained, and thus by taking various values of densities for sediments, crust and mantle, the relative size of mantle upwelling and basin-fill can be calculated. It is found that these two volumes are of a similar size, with the latter being perhaps larger.

Using this method, crustal extensions have been estimated for the basin cross-sections of Figs. 4.4, 4.7, and 4.10. The cross-sectional area of sediment is measured using a grid overlay, and an equal amount of mantle-upwelling assumed. Taking a crustal thickness of 40 km, the shortening which occurs when the "thinned" section of the crust is converted back

to a 40 km slab can be calculated, thus giving a rough estimate of the crustal extension. For each profile, the area of sediment deposited prior to cessation of faulting (i.e: pre-Eocene) has been taken. The results are given in Table 4.1. Individual estimates are probably rather unreliable, since on some profiles basement depth is only roughly estimated. However, compared between basins (see also the average estimates in Table 4.1) there are differences which are considered significant.

Crustal extensions are least in the Robe/Penola Trough (pre-Otway Group) and the Bass Basin. In both, the variation in extensions generally reflects the basin shape, decreasing towards the end(s). In the Gippsland Basin only the Strzelecki Group and the pre-Eocene Latrobe Group have been included in the calculations (thus neglecting later Latrobe Group sediments, which merely onlap basement), in order to estimate only fault-controlled extensions. The results suggest that the basin closes to the east, as earlier suggested (Section 4.1.4), but better basement control is needed to confirm this. Similarly only pre-Eocene sediments are considered in the Otway Basin, so that the extensions reflect the maximum rift-development before the onset of sea-floor spreading. Extensions are about twice as much, both expressed as percentages and as actual totals (the latter may be only about half the original extension, as the other half of the basin is probably now part of Antarctica). The general conclusion is that the extensions of close to 20% are characteristic of rifts which have subsequently developed into ocean basins.

The general relationship between rift-forming faults and older basement structures is illustrated in Fig. 4.17. Inspecting this figure,

TABLE 4.1: BASIN EXTENSIONS CALCULATED FROM CROSS-SECTIONS

(a) OTWAY (total pre-Eocene)

Profile	Length (km)	Area	Extension (km)	%Extension
1	80	282	14.10	21.39
2	100	305	15.25	17.99
3	110	330	16.53	17.68
4	120	430	21.52	21.35
5	140	447	22.36	19.00
6	120	420	21.00	21.21
7	110	344	17.21	18.54
8	120	337	16.83	16.31
9	100	294	14.68	17.20
10	100	413	20.69	26.08
11	100	322	16.14	19.24
12	100	333	16.67	20.00
13	80	158	7.92	10.98
14	100	265	13.28	15.31
15	50	155	7.75	16.97
16	60	143	7.17	13.56
AVERAGE OF 16 PROFILES			15.6	18.3

(b) OTWAY (Robe-Penola Trough pre-Otway Group only)

Profile	Length (km)	Area	Extension (km)	% Extension
1	80	174	8.70	12.20
2	90	171	8.55	10.56
3	90	106	5.3	6.25
4	90	62	1.03	3.57
5	50	46	2.30	4.82
AVERAGE OF 5 PROFILES			5.2	7.5

(c) BASS (total pre-Eocene)

Profile	Length (km)	Area	Extension (km)	% Extension
1	90	35	1.75	1.98
2	140	65	3.23	2.37
3	140	118	5.90	4.39
4	140	150	7.50	5.66
5	150	209	10.45	7.49
6	140	200	10.00	7.69
7	140	203	10.15	7.82
8	170	259	12.95	8.24
9	170	291	14.55	9.36
10	110	203	10.15	10.17
11	100	232	11.60	13.12
12	50	34	1.70	3.92

AVERAGE OF 12 PROFILES

8.3

6.9

(d) GIPPSLAND (total pre-Eocene)

Profile	Length (km)	Area	Extension (km)	% Extension
1	90	53	2.67	3.05
2	90	81	4.07	4.73
3	90	110	5.49	6.50
4	110	196	9.8	9.78
5	140	282	14.2	11.28
6	130	351	17.5	15.55
7	130	361	18.0	16.07
8	120	302	15.0	14.29
9	150	320	16.0	12.12
10	50	78	3.9	8.46

AVERAGE OF 10 PROFILES

9.6

11.4

it is apparent that there is in fact very little obvious relationship between the two sets of structures on a regional scale. The main rift axes (heavy dots) cut right across the older trends, for example west of Tasmania. Although the main Australia/Antarctica rift turns southwards, it appears neither to be controlled by, nor tend to become parallel to, older structures. The Tasman Sea margin roughly parallels Paleozoic trends on Fig. 4.17, but further north, off the New South Wales coast, it too cuts across the Paleozoic structure (see Fig. 2.5). On a more local scale, a similar non-relationship is evident, but, since little is known of the basement structure beneath the basins, this cannot be so clearly demonstrated. The Bass Basin is roughly aligned parallel to Paleozoic trends projected from Victoria and Tasmania, but the Gippsland Basin, especially its northern margin, cuts right across similar trends. The distribution of granites may also have some influence, for example along the Bassian rise, but again this cannot be demonstrated as their subcrop limits are not known.

The general picture which emerges from the structural analysis is of an essentially homogeneous basement, which, at least near the surface, behaves as a brittle slab in the applied tensional regime. The development of rift-forming faults is neither markedly influenced by pre-existing structures, nor related to the direction of later sea-floor spreading. The fault-pattern should therefore show some simple relationship to the regional principal tension established, influenced only by any variation in the orientation of this stress.

A generalised analysis of the local stress-distributions is shown in Fig. 4.18. The inset shows the principal tension (N35°E) and the trends of related normal and shear faults, and second-order faults, and folds. On the map, possible local stress-distributions related to this major direction are indicated by strain ellipses. Many of the Otway and West Tasmanian faults are interpreted as normal faults, and others as second-order faults associated with a sinistral shear. In the Bass and Gippsland Basins, *en echelon* sets of faults can be related to dextral shear, and an overall dextral shear probably controls the Bass Basin development. To the north, in the Otway Ranges and South Gippsland Highlands, fold axes have a fairly consistent orientation, which can be related to a similar dextral shear. The magnitude of these movements can be judged from the scale (maximum extensions range up to about 15-20 km in Bass and Gippsland) and is relatively small compared to the size of the basins. The total shift between Tasmania and mainland Australia is thus only about 10-20 km. All of the major structures can thus be related to the pre-Eocene regional stress pattern - i.e: prior to sea-floor spreading between Australia and Antarctica. Later movements are largely confined to the Gippsland Basin, and can be related to re-activation of pre-existing Jurassic to Early Eocene structures. Thus again the conclusion is that rift formation seems to be independent of any control by sea-floor spreading and related transform faults.

In the following three Sections, the geological evolution of south-east Australia is considered in the context of the regional tectonic and structural framework outlined above. Three arbitrary divisions correspond to the early rifting stage, the Tasman Sea opening and the Australia/Antarctica separation, but the sequence of events is

really continuous. Geological data are integrated with the regional tectonic developments at each stage, and an evolutionary synthesis developed which is compatible with both.

4.2.3: JURASSIC AND LOWER CRETACEOUS - RIFT VALLEY DEVELOPMENT

During the Late Paleozoic and Early Mesozoic, there is little evidence of any tectonic activity near south-east Australia. 1000 km to the east, Carboniferous to Early Permian subduction and volcanism was followed by sedimentation along the Gondwanaland margin (Chapter 3.2). Elsewhere relatively thin Permian and Triassic platform cover was deposited across large areas of the Gondwana craton (Chapter 2.3.3; Fig. 2.11), and some remnants of these sediments occur in Tasmania, the Murray Basin and elsewhere.

The earliest sediments recognised as being deposited in rifts are of Jurassic age, occurring in the Polda Trough (Smith and Kamerling, 1969; Harris, 1964), the Penola Trough (Casterton WCR; Cundill *in* Parkin, 1969, p. 150) and probably at the base of the succession in parts of the Otway and Gippsland Basins. During the Lower Cretaceous, much greater volumes of sediment were deposited in the extensive rift system developing south of Australia.

The structural framework of Jurassic and Lower Cretaceous deposition is shown in Fig. 4.19. In the areas of the Otway, Bass and Gippsland Basins, faults and isopachs can be drawn fairly accurately, but elsewhere the structure is only generalised (on the Australian side) or inferred (on the Antarctic side). The major rift varies from 50 to 150 km wide. It probably connected with the Indian ocean in the west, developing progressively eastwards during the Jurassic. Several prominent "splays" or subsidiary rifts occur along the northern side, and seem to be related

to changes in the trend of the main rift. The Polda Trough cuts into the Proterozoic Gawler Block. The extension of its southern flank is probably the Ceduna Plateau. The Robe-Penola Trough is a similar splay, bounded to the south by the Kalangadoo High and the Beachport Plateau. Further east the rift system becomes more complex. Originally it may have splayed north-east into the Torquay Embayment. Other rift-basins (Gippsland, Bass) also began to form during this stage. No data are available as to the termination or continuation of the rift system in the area to the east of Australia, but there is no evidence to suggest that it broke through into the Pacific Ocean, along the margin of which subduction was occurring during the Upper Jurassic to Lower Cretaceous (Chapter 3.2.3).

The Jurassic and Lower Cretaceous evolution of the area is illustrated by two schematic block-diagrams (Figs. 4.20, 4.22), and accompanying paleogeographic maps (Figs. 4.21, 4.23) showing pre-Otway Group and Otway Group (and equivalents) respectively. In the initial stage of development (Fig. 4.20) the rift system reached south-east Australia in the Jurassic. The rift may have forked west of King Island, with the major branch running north-east into the Torquay Embayment. A series of *en echelon* faults then dissipated the regional tension eastwards across Victoria, initiating the "Strzelecki Basin". These structures are schematically outlined in Fig. 4.20 (lower). During this stage, Tasmania is regarded as part of Antarctica, with faulting occurring mainly to the north.

The early paleogeography (Fig. 4.21) is reconstructed from only a few pre-Otway Group occurrences (Section 4.1). However, these indicate essentially deposition of clastics derived from the rift margins. The

Tyers Conglomerate in Gippsland (Philip, 1958) is probably typical of the earliest sediments. Igneous activity is evident in the Penola Trough and elsewhere (see Section 4.2.6). The Penola Trough seems to have been the earliest rift in the Gambier Embayment, but later the main rift valley developed south of the Kalangadoo High (c.f: Rochow, 1971a, p. 314). As the trough filled, thin coal measures were formed.

During the rest of the Lower Cretaceous, an essentially similar tectonic regime prevailed (Fig. 4.22). The main rift through the Otway Basin continued to widen and deepen. To the east the *en echelon* fault systems developed further, and the Torquay Embayment and "Strzelecki Basin" deepened. However, it seems that this fault system, once established, prevented further eastward extension of the main rift. Increasing displacement was thus accommodated by initiation of the Bass Basin, dissipating the overall shear between Tasmania and Victoria, and by propagation of the Otway Basin to the south-east, along the western side of Tasmania (Fig. 4.22 lower). At the same time, the Tasman Sea may have begun to form as a rift, and the Gippsland Basin may thus have terminated to the east in the incipient Tasman Sea, rather than extending onto the Lord Howe Rise.

Lower Cretaceous paleogeography (Fig. 4.23) during this time reflects the extensive rift-developemnt. Thick sequences of clastics, comprising arkoses, greywackes, silts and muds, were deposited in the Otway, Gippsland (and most probably Bass) Basins. In the Otway Basin there is evidence of provenance from both northern and southern margins (e.g: Otway ER-68 SS, p. 4). Coal measures in some areas probably indicate deltaic conditions near the basin margins.

To summarise, the Jurassic and Lower Cretaceous saw the development of an extensive rift valley and basin system between Australia and Antarctica. The structural framework controlling the later evolution was essentially established during this time, but as previously indicated (Section 4.22) there is no clear evidence to suggest that basement structures controlled or influenced the rift evolution. Most of the structures can be interpreted as related to a principal tensional stress at N35°E acting across the region. Sedimentation in the rifts was entirely non-marine, consisting predominantly of clastics derived from erosion of the basement along the rift margins, with some volcanics and coal seams.

4.2.4: *UPPER CRETACEOUS TO EARLY EOCENE*

The general Upper Cretaceous and Paleocene structural framework is shown in Fig. 4.24. Isopachs are generalised along the southern margin, and not shown west of the Otway Basin, where however a considerable thickness of Upper Cretaceous is probably present. In comparison with Fig. 4.19, the rift is now wider (up to 250 km) and is opening along the western side of Tasmania. Upper Cretaceous sediments are absent in the Otway Ranges and South Gippsland Highlands, but the Bass Basin reached its maximum development during this time, extending onshore as the Tamar Graben by the Paleocene. Generally the structural pattern is similar, but the basins more extensive. By the mid-Upper Cretaceous, oceanic crust was present in the Tasman Sea to the east. The Upper Cretaceous to Early Eocene evolution is illustrated by two block-diagrams (Figs. 4.25, 4.27), and accompanying paleogeographic maps (Figs. 4.26, 4.28) showing the Upper Cretaceous and Paleocene to Early Eocene respectively.

In the Upper Cretaceous, development of the main rift system continued, with further downfaulting in the Otway Basin. With increasing

relative movement between Tasmania and Antarctica, the dextral shear-stress across southern Victoria increased, and related folding (see inset, Fig. 4.18) probably began to occur, producing the NE-trending anticlinal structures in the Lower Cretaceous sediments of the Otway Ranges and South Gippsland Highlands, and consequent cessation of deposition in these areas (Fig. 4.25 lower). Thus the regional tension was now dissipated across the Bass Strait area to the south, with further deepening of the Bass Basin, and its extension into northern Tasmania. Development of several other NW-SE trending horst and graben structures in Tasmania also occurred at this time (c.f: Banks, *in* Spry and Banks, 1962, pp. 241-2, fig. 39).

Throughout this period Tasmania was becoming increasingly detached from Antarctica, and by the Paleocene, separation was almost complete. By this time the Otway rift had extended south along the western side of Tasmania, with a splay forming the Macquarie Graben (Fig. 4.25). Some of the faulting along this part of the rift may have resulted from a sinistral shear related to the $N35^{\circ}E$ principal tension, but it is also probable that this regional tension swung around towards the direction of incipient sea-floor spreading during the final stages of rifting. The most likely effect of this would probably have been reinforcement of favourably-oriented earlier structures, rather than development of new faults.

To the east, the Tasman Sea was also evolving during the Upper Cretaceous. Its margin is only poorly mapped seismically, and only a few down-to-the-east faults are known (East Gippsland/Tasman/Bass Strait SS; East Tasmania T69B SS). Thus the early evolution of the Tasman Sea margin cannot be studied in detail. The present shelf is relatively

narrow, but this may in part reflect a paucity of sediment supply. However, it seems likely that the initial rift preceding sea-floor spreading was of considerably smaller dimensions than the Otway rift. The oceanic crust just east of Bass Strait is at least 80 MY old (Hayes and Ringis, 1972) and so by the Late Upper Cretaceous, the eastern margin of Australia/Antarctica was being carried away from the Tasman Sea spreading axis, which played no further role in the regional evolution.

Upper Cretaceous paleogeography (Fig. 4.26) reflects the continuing rift-development in the west. The basal unit of the Sherbrook Group (Fig. 4.5) is a transitional series of sands and grits (Waare Formation) probably composed of re-worked Otway Group sediments. This is followed by a marine transgressive sequence (Flaxmans Formation, Paaratte Formation) which appears to indicate a slow influx of the sea from the west. This seaway developed as a long inlet of the Indian Ocean, and considerably influenced the later stratigraphy. The Otway Basin section shows a series of marine transgressive/regressive cycles (c.f: Bock and Glenie, 1965; Taylor, 1971). The Bass Basin remained unconnected with this sea, and filled with a non-marine deltaic series of sands and shales, with coals indicating shallow water conditions generally.

In the east, the presence of the Tasman Sea profoundly influenced the Gippsland Basin development. The earlier "Strzelecki Basin" was probably entirely bounded by continental crust, like the Bass Basin, but as the eastern part was pulled away, the basin opened to the Tasman Sea, and a large delta system began to build out onto the ocean floor. The Upper

Cretaceous Latrobe Group (Fig. 4.11) is entirely non-marine where drilled, but marine evidence might be expected in the extreme east. Little is known of the depositional conditions elsewhere along the Tasman Sea margin, but in Tasmania the Derwent River delta may have begun to form about this time.

In the Paleocene to Early Eocene (Figs. 4.27, 4.28) the Otway rift reached its maximum development. At this stage the centre was probably deeply submerged, and the paleogeography of the Australian and Antarctic margins might then have begun to show increasing diversity. True sea-floor spreading began near the base of the Eocene (55 MY ago), and a mid-ocean ridge evolved in the centre of the rift. The course of this ridge was thus controlled by the *earlier* rift, which turns south along the Tasmanian margin. Plate-tectonic geometry requires that spreading ridges be aligned roughly normal to the vector of motion between separating plates. Hence such a major bend in the rift would impose a constraint on ridge development, leading to its offset by a series of transform faults. Initially the ridge segments may have been very small, and hence the magnetic pattern in the oldest oceanic crust would only be detected by close profiling.

The arm of the south-east Indian Ocean broke through into the Bass Basin in the Paleocene, and marine deposition progressively took over from the earlier fluvio-deltaic conditions. In the Gippsland Basin, the Latrobe Delta continued to build out into the Tasman Sea, and also to transgress over the earlier basin flanks. The Tuna/Flounder and Marlin Channels were major distributaries which incised into the delta top. The earliest known marine influence occurs in these channels (Flounder and Turrum Formations),

as the eastern sea encroached and mixed with clastic sediments.

In summary, the Upper Cretaceous and earliest Tertiary was a period of increasing rift valley and basin development, largely within the structural framework established earlier. As the main rift deepened, the Indian Ocean entered from the west. At the same time, the Tasman Sea was opening and the Gippsland Basin became a large delta complex. The insertion of oceanic crust between Australia and Antarctica also marks the end of major fault-movements in the region. The previously noted widespread Eocene unconformity (Section 4.1) reflects the end of the structural evolution of south-east Australia, except for later local rejuvenation of some faults (see below).

4.2.5: *LATE EOCENE TO RECENT*

The extent and thickness of Upper Eocene and younger sediments is shown in Fig. 4.29. Up to 2000 m of sediments occur in the Latrobe Delta of Gippsland and 1000 m in the centre of the Bass Basin. In the Otway Basin prograding sediments are up to 1000 m thick on the edge of the continental shelf, but wedge out southwards. Thin Tertiary sediments also occupy an extensive area in the Murray Basin. By this time Antarctica was shifting away to the south, and little is known of depositional conditions, which however would have become increasingly different to Australia as the Antarctic climate cooled and the present ice-cap developed. There is little evidence for further major structural development after the Eocene. Fault movements of up to a hundred metres or more are recorded, and the relatively low, but significant, seismic activity (Fig. 4.13) indicates continuing mild tectonism. In the onshore Gippsland Basin, Hocking (1970, fig. 5) shows that monoclinal folding of

sediments as young as Pleistocene is associated with continuing movement of older structures.

These movements are regarded as due to eperiogenic readjustments *within* the Australian plate after its separation from Antarctica. The concept of an Upper Tertiary dextral shear in the Gippsland Basin (e.g. Richards and Hopkins, 1969), and the lack of any unequivocal supporting geological evidence, has already been mentioned in Section 4.1.4. Once the south-east Australian region was separated from the Australian/Antarctic plate boundary by oceanic crust, any significant *horizontal* displacements could only occur if they linked with a plate boundary through oceanic crust. There is no evidence to suggest eastward extension of an active Tertiary fault eastwards into the older Tasman Sea, or its westward extension and connection with the south-east Indian Ocean. Such a postulated dextral shear cannot be precluded on this lack of evidence alone, but it has also been shown that the structural framework of the Gippsland Basin can be accounted for largely within the Upper Mesozoic to Early Eocene stress regime, and hence there is no requirement for an Upper Tertiary dextral shear to explain the Gippsland structures. The conclusions of Hocking (1972) that Cenozoic structures are related to rejuvenation of older blocks is therefore preferred, and it is more consistent with the overall hypothesis presented here.

A block diagram (Fig. 4.30) shows the Tertiary setting of south-east Australia. The early transform-fault system west of Tasmania probably simplified into a few large transforms soon after sea-floor spreading had begun (compare Fig. 4.27), but these had no controlling influence on the earlier structural evolution of the Tasmanian

margin. The South Tasman Rise probably remained partially attached to Australia until the last, becoming rifted from Tasmania, but then the ridge developed to the south. Tasmania and the South Tasman Ridge together form a long, tapering peninsular south of Australia (see Fig. 4.13). This continental crust may have thinned during separation, with consequential isostatic uplift, perhaps explaining why Precambrian rocks are now exposed in western Tasmania, whereas the Lower Paleozoic still covers any older rocks in Victoria. Preliminary crustal thickness estimates (Johnson, 1972; J. Knight, pers. comm.) indicate a thinner crust under Tasmania, particularly in the west, than in Victoria, supporting this suggestion. The submergence of the South Tasman Rise suggests it may be even thinner crust.

Late Eocene to Miocene paleogeography is summarised in Fig. 4.31. The quiet tectonic conditions following complete detachment of Australia and Antarctica are reflected by the predominance of mudstones, siltstones and limestones. The south-east Indian Ocean spread into the Bass Basin, and in the Miocene breached the Bassian Rise. Shallow seas also extend well inland into the Murray Basin, and also along the north-western margin of Tasmania. The only major clastic deposition was in the Latrobe (and possibly Derwent) deltas. In the Gippsland Basin, the sea transgressed westwards across the Latrobe delta (see Fig. 4.11), and the diachronous Lakes Entrance Formation was deposited over the topographically uneven delta-surface, later to form the cap-rock over the oil-bearing structures. Non-marine deltaic conditions became confined to the onshore area of the basin, and during this period the extensive brown coals of the Latrobe Valley were laid down. Channel-cutting continued intermittently, reflecting epeirogenic adjustments and perhaps

sea-level changes. Oil and gas migration and accumulation must also have occurred during the Upper Tertiary, the hydrocarbons probably being derived from within the Latrobe Valley Group.

4.2.6: IGNEOUS ACTIVITY DURING BASIN EVOLUTION

Mesozoic and Cenozoic igneous activity in south-east Australia and the adjacent part of Antarctica is summarised in Fig. 4.32. Three broad types of igneous activity can be distinguished: (a) Jurassic dolerites; (b) Upper Mesozoic igneous rocks; (c) Cenozoic basalts. These three types exhibit different relationships both in space and time to the basin development.

Jurassic dolerites occur in Tasmania and Antarctica. In Tasmania they form differentiated tholeiitic sills and feeder dykes, generally within fairly flat-lying Permo-Triassic sediments (c.f: Spry and Banks, 1962, pp. 266-70). In Victoria Land, Antarctica they occur as tholeiitic sills and as flood basalts.* In both areas they are dated as Early Jurassic. The dolerites have no obvious spatial relationship to the rifts, and pre-date the oldest rift sediments. Thus their connection with continental break-up appears slight, but they may herald the earliest signs of tectonism in the region.

The "Upper Mesozoic igneous rocks" are known from only a few localities in south-east Australia. A number of different rock-types are included in this category. Sutherland (1972) describes shoshonitic rocks from several localities in Tasmania, where they occur as small intrusive bodies of Lower to Middle Cretaceous age. Rocks of similar age have been encountered in wells in Victoria and Bass Strait (see list on Fig. 4.32, and refer appropriate WCR's). Some of these are

*c.f: Gair *et al.*, 1970

briefly described by Hocking (*in* Hawkesdale WCR), but the samples are very small and no analyses have been made. Rock-types appear to include olivine basalts, trachytes, tuffs and indeterminate, very altered volcanics. The known occurrences of these Upper Mesozoic rocks are limited, but seem to have a fairly close spatial relationship to the early rifts. The African rift valley system has extensive associated volcanism, and it may be that the few known occurrences in south-east Australia also reflect a volcanic phase in the early rift evolution, now largely obscured by later sedimentation along the continental margin.

The Cenozoic basalts are widespread, in a province extending the length of eastern Australia (c.f: Geological Society of Australia, 1971) and into Antarctica (c.f: Gair *et al.*, 1970). Their distribution is not indicative of any relationship to the rift history, and their ages have no obvious correlation with the onset of sea-floor spreading. They erupted well away from the plate margins, and hence are regarded as intra-continental basalts, perhaps indicative of "mantle hot spots", but otherwise with little tectonic significance.

4.2.7: EVALUATION AND CONCLUSIONS

The synthesis outlined in this Chapter has implicitly assumed the validity of the sea-floor spreading hypothesis, in that it is based on the regional model constructed in Chapter 2. The geological data all appear consistent with this assumption, but it is also possible to specifically evaluate the sea-floor spreading hypothesis against the known geological data, as a further test. To do this, I have taken the Otway Basin, which is a "true" continental margin in that there is a transition from continental to oceanic crust across the basin, and used the time-space

analysis method developed in Appendix A (see especially Sections A.5, A.6.1). The basic scheme as applied to the Otway Basin is shown as a flow-chart in Fig. 4.33 (compare with Figs. A.6, A.7). The data used in construction of the "actual time-space plot" (Fig. 4.34) is derived from the literature as reviewed in Chapter 4.1.2., in which the known geology of the Otway Basin is described without reference to any tectonic interpretation. The "model time-space plot" (Fig. 4.35) is based on a model for continental break-up beginning with a rift valley and proceeding to sea-floor spreading. If the two plots are significantly different, then only the *model* can be modified. Fig. 4.35 thus represents a model considered to be acceptably similar to Fig. 4.34.

The time-space plot of the Otway Basin (Fig. 4.34) is based on a cross-section through the Gambier Embayment of South Australia, roughly normal to the strike of the continental margin, and passing through the Penola Trough. The two lower insets show the location and generalised geology (see also Figs. 4.3, 4.4, 4.5 for detail). The time-space plot is drawn roughly to scale. At the base are the "Casterton Beds" and Pretty Hill Sandstone, which include some of the Upper Mesozoic igneous rocks described in Section 4.2.6., and overlie basement or thin permian sediments. During this phase the Kalangadoo High and Penola Trough remained as distinct features, but later the Otway Group was deposited right across them. Some Otway Group sediments appear to have been derived from the south (Otway ER-68 SS, p. 4). The Sherbrook Group represents a longitudinal marine incursion into the basin. All these units thin seawards, and above the Eocene unconformity which also marks the end of fault movements, prograding wedges of sediments are forming. Confirmation of the age of the adjacent oceanic crust awaits the results of JOIDES Leg 29, but

magnetic data suggests an Eocene age. The Upper Tertiary is a period of quiet deposition, with continuation of a succession of transgressive/regressive cycles.

The model time-space plot is based on four cross-sections showing the postulated evolution of a rift valley, corresponding roughly to the Rhine Graben, African Rift, Red Sea and Gulf of Aden (see Griffiths, 1971b, fig. 1). A horst block and minor rift is introduced in A and B, to simulate the Penola Trough, and subsides in C. The time-space plot illustrates the evolutionary sequence. At the bottom, a roughly symmetrical rift valley develops, with another minor rift to the right. Clastic sediments fill the rifts, and fairly extensive volcanism occurs (A, B). Later this diminishes, and the sea enters along the major rift (C). Finally a mid-ocean ridge forms and one side of the rift is carried rapidly away. Each half (only one is shown) is then a separate "Atlantic-type" continental margin (D).

The two plots can now be visually compared, and their compatibility assessed. My own conclusion is that an acceptable degree of similarity exists, and hence that the sea-floor spreading hypothesis leads to a satisfactory synthesis of the evolution of the Otway Basin section analysed. To completely test the synthesis, more sections could similarly be treated. It is also noted that the model *predicts* the data which should occupy the gaps in Fig. 4.34 on the seaward side. Thus the sediments in this area should be derived largely from the south. Much of this area should now lie on the Antarctic margin, where a broadly similar stratigraphy until the Eocene is therefore predicted. Following separation, climatological controls would become far more significant, with possibility of very divergent stratigraphic successions in the rest of

the Cenozoic. Such predictions may have some relevance to the search for oil, but it must be remembered that the gross structural framework is also important. Thus on the Antarctic margin there is no partially-detached "sub-plate" such as Tasmania, and therefore no equivalent of the Bass or Gippsland Basins is expected.

Finally the history of the Bass and Gippsland Basins can be tested for compatibility with the synthesis. Time-space plots could again be used, but here I have summarised the data in terms of time-relationships only (Fig. 4.36). The first column gives the sea-floor spreading history, shown above to be compatible with the Otway Basin history (second column). Data from the Bass and Gippsland Basins, and igneous activity, are taken from Sections 4.1.3, 4.1.4 and 4.2.6 respectively. Detailed inspection of Fig. 4.36 will show that the general relationships between faulting, igneous activity and depositional conditions correspond fairly closely with phases in the tectonic history, and generally accord with the synthesis of south-east Australia presented in this Chapter.

APPENDIX A

A DISCUSSION OF THEORETICAL ASPECTS AND OBJECTIVE METHODS OF
TECTONIC INTERPRETATION

Tectonics is essentially the study of the structural features of the earth, and the causes and forces which have produced these. The ultimate aim of a regional tectonic study is to present a dynamic synthesis of the geological history of that region. The study of tectonics thus encompasses all aspects of earth science, in the endeavour to produce the most comprehensive interpretation of any given data.

Most regional tectonic studies either explicitly or implicitly assume the validity of a tectonic hypothesis as a basis for data interpretation. As a general observation, there is little positive attempt in the literature to distinguish hypothesis from geological fact, and as a result most tectonic studies lack the objectiveness which the field geologist strives to attain. It is with this need for an objective approach to tectonic interpretation in mind that, as I have studied the regional geology of the south-west Pacific, I have also tried to develop new techniques for the handling of geological data. This Appendix is an attempt to discuss these methods in a more general context, in the hope that they may be further developed in the future, and provide the basis for a more objective approach in tectonics.

A.1: GENERAL PRINCIPLES

The process of tectonic interpretation of geological data can be considered in two stages:

- (i) Tectonic analysis - collection, processing and presentation of data;
- (ii) Tectonic synthesis - interpretation of the results of (i) in terms of a tectonic hypothesis.

The field geologist is primarily concerned with tectonic analysis, whereas tectonic synthesis is generally considered to be an academic exercise. Objective methods are (or should be) the aim of both field geologist and academic alike, but the latter too often either consciously or otherwise applies his own preferred tectonic hypothesis to geological data in a purely subjective manner. There should however be no difference in approach, as tectonic synthesis is only the next logical step beyond tectonic analysis. A tectonic synthesis can be thought of as a *dynamic model*, based on an explicitly stated hypothesis, which simulates the available geological data. Such a synthesis has two important characteristics - it can be readily and objectively evaluated, and it is predictive.

A.1.1: TECTONIC ANALYSIS

Tectonic analysis is in fact little more than the normal sequence followed by any geologist studying and writing a report on an area. The aim is to produce an account of the geological features observed in the field and laboratory, and to make deductions about the geological history. These observations and deductions form

the basis upon which any subsequent tectonic interpretation must rest. Hence it is vital that tectonic analysis should be as objective as possible, and should not incorporate any notions based upon particular hypotheses.

Collection of geological data generally begins in the field. Numerous descriptive parameters are used. Some, such as those which define position (e.g: latitude, longitude, elevation) are quantitative numerical parameters, but most geological parameters, such as field descriptions of rocks, are more subjective and qualitative. Structural information is an exception, as it concerns measured elements. Laboratory study of geological data tends to reduce the subjective elements of field descriptions to a degree, as numerical schemes of classification can be erected. However, many results are still presented in non-numerical form (e.g: rock names, fossil names) which then cannot be readily duplicated by other workers. Truly objective descriptions are thus rarely achieved in geology, and this must be kept in mind when any data processing techniques are used. Development of numerical data recording techniques and computer storage and handling of data would seem to be the logical way to reduce the subjective elements inherent in most descriptions.

The next stage in tectonic analysis is the initial interpretation of data. The aim is to determine, from the raw data available, the history of events recorded by the data, without at this stage considering the causes or processes responsible for those events. Examples include such familiar techniques as determination of paleocurrent directions, sequences of structural

and metamorphic events, and so forth. Some of this interpretation involves acceptance of particular theories, such as those based on experimental work, but at no stage must tectonic hypotheses be invoked. At this stage the concern is only to derive the maximum factual information from the data available.

The final stage involves a greater measure of interpretation, and extrapolation of data in space and time. The result of this should be a geological history based on the data, or in other words a record of the geological events which have occurred in a particular region. This is presented both as diagrams (maps, sections, etc.) and as narrative as appropriate. Together these should give an account of the geological history, but as yet the causes controlling the evolutionary sequence have not been considered, and tectonic hypotheses have not been introduced.

A.1.2: *TECTONIC SYNTHESIS*

The relationship between tectonic synthesis and analysis is illustrated most clearly by using a *flow chart* (Fig. A.1). The construction of this and other flow charts is itself a step towards a more objective approach in tectonics. Flow charts provide a clear picture of the steps followed in a multi-stage operation, and allow careful evaluation of any scheme. They are constructed in a similar manner to those used in computing. Each step must follow logically from the previous one, perhaps with input or output of data, or involving comparisons and other decisions. Thus any illogical steps, which might be easily missed or glossed over in a narrative description, become obvious. The flow charts themselves

are largely self-explanatory, and several different examples are given in this thesis.

Figure A.1 thus outlines the fundamental basis of tectonic synthesis. The stages of tectonic analysis follow those discussed above (Section A.1.1). In order to construct a synthesis of the geological data, it is necessary to adopt a particular hypothesis (e.g: the "tectogene" model; plate tectonics; etc.), which has been derived by inductive reasoning from general geological observations. This hypothesis then allows an *actualistic model* to be constructed, such that it simulates the real data, and hence the evolution in space and time of the region studied. When the parameters of the model have been adjusted so that it most closely simulates the real situation, it can be adopted as a tectonic synthesis of the region.

This synthesis must then be *critically evaluated* against the basic geological data. In other words - does it adequately account for and satisfactorily explain *all* aspects of the data? The evaluation is carried out by feed-back from the synthesis to the basic data, carefully checking their mutual consistency. If any inconsistency is found, then either (a) the basic data is inaccurately recorded, or (b) the synthesis is unsatisfactory. The first possibility is readily checked by re-examination of the basic input data for accuracy. If the synthesis is still inadequate, then either the actualistic model must be further modified, or the basic hypothesis itself modified or rejected. The evaluation procedure can be continued until an acceptable synthesis is made, but at no stage is there provision for rejection or alteration (other than correction) of any of the real geological data.

The scheme also has one other important characteristic, which is only evident in objectively based methods of tectonic interpretation. The tectonic synthesis is *predictive*. Thus, once constructed, the synthesis will point to and predict new data which can be collected, thus providing the basis for an independent evaluation of the synthesis by other workers. This is perhaps the ultimate test of a tectonic synthesis - how successfully can it predict new data? Once confident predictions can be made, as now seems possible in the light of plate tectonics, then tectonics will be established not just as an academic exercise, but as a valuable tool in geological exploration.

Unfortunately there are still many examples of subjective, illogical approaches to tectonic interpretation, most of which would be obvious if they were carefully analysed using flow charts as outlined above. As an example, I have shown a simple general case of a tectonic synthesis based on an incorrect approach (Fig. A.2) which is extremely common in the literature. In this case the tectonic hypothesis has been used to construct an actualistic model, as before, but then the model has been used during the tectonic analysis, rather than in interpretation of the results of this analysis.

A common example of this approach is the case in which a plate tectonic model is postulated for an area, and then geological data which "fits" the model is sought. If the model is obviously wrong, it will be discarded, but the danger is that only the data which is consistent with the model will be considered. Thus the

basic stages in interpretation of geological data, which should be as objective as possible, are now dependent upon the plate-tectonic hypothesis. There is no way out of this circle if the model is adhered to, and if it is modified or changed, much of the basic interpretation must also be reconsidered. The system has no inbuilt "feed-back" tests as in Fig. A.1, and as a consequence will generally fail to fully utilise or be consistent with all the available geological data. In addition the predictive value of such a hypothesis is likely to be small, and instead only further data are sought which will confirm the synthesis, rather than critically test its validity.

It is therefore necessary to take great care not to use such theoretically unsound schemes as a basis for tectonic synthesis. Similarly care is needed in critical evaluation of tectonic syntheses proposed in the literature, to determine whether such unsound schemes of analysis have been used. This is often difficult, as without full knowledge of the regional geology, "feed-back" tests cannot be applied by the reader. If however such tests can be carried out, and fail, then a very careful evaluation of the synthesis should be undertaken.

As a particularly obvious example, I would mention the concept of "geosynclines". This area of geology has involved a great deal of debate, and many definitions and classifications have been advanced, with little agreement being reached. The concept of geosynclines has of course proved a very useful approach in geology, but it has also led to much confusion and argument. This is probably because it has been introduced at the wrong point

in tectonic synthesis (see Fig. A.3).

The concept of geosynclines properly belongs to the realm of tectonic hypothesis and actualistic models, and should therefore not be introduced until *after* tectonic analysis has been concluded. It is, however, generally but incorrectly introduced *during* analysis, and this leads to subsequent confusion. The geological history thus acquires an inherent subjective content based on tectonic hypothesis, and this effectively blocks any feed-back evaluation of a later tectonic synthesis (Fig. A.3). This error, once perpetrated, is often not noticed. Thus many geological reports present conclusions using schemes of geosynclinal nomenclature. The unwary geologist, attempting a tectonic synthesis from the wrong angle, is then in effect trying to use plate tectonics (or other hypotheses) to synthesise data which already involve a synthetic component.

Thus I would propose that the use of geosynclinal nomenclature is both unnecessary and no longer has any valid place in tectonic interpretation. Whilst this approach has had its merits in the past, as tectonic ideas were being formulated, I would urge that it now be abandoned, in favour of the more objective techniques of assembling data for tectonic synthesis, and for presentation of the synthesis, which I shall describe later.

If the basic scheme of tectonic analysis which I have outlined (Fig. A.1) has been followed rigorously, then, when a tectonic synthesis has been made, it can be exhaustively evaluated using the feed-back technique. New data will also continue to become available. These are easily handled by first putting them through the three phases of

tectonic analysis, then building on or modifying the synthesis by changing the model parameters. If however an invalid scheme has been used (e.g: Fig. A.2), the tendency is to incorporate these new data into the already established synthesis. The new data are now being interpreted in terms of the synthesis, by-passing the objective stages of tectonic analysis (see Fig. A.2). The implications of these general principles of tectonic analysis and synthesis are thus extremely important, and should be constantly kept in mind.

I consider that the basic scheme that I have outlined in Figure A.1 to be the fundamental basis of tectonic synthesis, and the more elaborate schemes which I shall develop later all have the one essential characteristic that tectonic hypothesis is not introduced until *after* the analysis has been carried out. Firm adherence to this principle allows maximum use of the plate-tectonic hypothesis to be made in geology.

A.2: RELATIONSHIPS OF DESCRIPTIVE PARAMETERS AND DATA PLOTS

A geological history can be presented using many different types of data plots, such as maps, cross-sections, isopach maps, etc. Because these plots are all based on the same original data, they must all be related in some way. In order to examine these inter-relationships, and to develop possible new plots, I have outlined a simple theoretical approach which shows clearly how basic data can be objectively recorded, and subsequently used to construct various data plots.

The basis of this approach is a scheme for recording geological data using five parameters. Consider a volume containing geological data. The variation in geological data within the given volume can be recorded by erecting a space reference frame based on three mutually perpendicular axes, and then for each point within the volume specifying its position and the geological information at that point. The age of the rock at any point, although it may be regarded as "geological information", is more conveniently handled by specifying it as a separate parameter.

As a simple example I will consider a sedimentary accumulation at a continental margin, which has distinct directional characteristics. A line parallel to the continent-ocean boundary is referred to as the "strike" of the margin. The basis of applying the descriptive scheme is then to record the age (T) and specified geological data (F) at points $P_1, P_2, P_3, \dots, P_n$ within the sedimentary accumulation at this margin. For the moment I will assume that the strike of the margin being analysed is a straight line, to simplify the discussion. The parameters used in data recording are then as follows:-

- (i) Space Co-ordinates (X,Y,Z).

Three mutually perpendicular reference axes

(X,Y,Z) are chosen such that they intersect at origin

0 (x_0, y_0, z_0). The orientation of these axes is defined thus:-

X is horizontal and perpendicular to the strike of the margin, Y is horizontal and parallel to the strike of the margin, Z is vertical (depth), the distance along it being measured downwards from a datum X-Y plane z_0 which is taken at sea-level.

This X-Y-Z reference frame can then be used to specify the position of any point within the continental margin. It can also be used to uniquely describe discrete planes and surfaces of geological significance, such as bedding-planes, unconformities, faults and the form of igneous bodies.

(ii) Time (T).

This parameter records the time of deposition of the sediment at each point. It can also be used for any other features with time-connotation, such as time of post-depositional modification, metamorphism, igneous intrusion, etc. In the real case it is deduced from geological evidence, and therefore as recorded at any point will represent the best age determination available at that point. Any surface with constant T-value (referred to as a T(t)-surface will be a time-stratigraphic surface.

(iii) Rock Factor (F).

This is the most complex parameter. It is a compound factor which records previously specified geological information from each point. It can therefore consist of many component parts, which in later data-processing can be retrieved separately or in combination as required. The F-factor should ideally contain sufficient data to (a) differentiate between rock-stratigraphic units (beds, formations, etc.) and (b) record lateral (i.e: facies)

variations within each unit. Any properties of a rock can be used as long as they can be expressed numerically. The literature contains many examples of suitable descriptive techniques and classifications. Three special surfaces are defined:-

$F(s)$ is a rock-stratigraphic surface (e.g: a bedding plane);

$F(fm)$ is used to denote that variation of F which defines a mappable lithostratigraphic unit (e.g: bed, formation), and as such specifies an upper and lower boundary surface;

$F(r)$ is used to denote internal lateral variation (facies change) within a specified lithostratigraphic unit $F(fm)$.

A complete three-dimensional set of points, with X , Y , Z , T , F recorded at each point, thus constitutes an objective description of part of a continental margin. In reality, a finite number of points must be used, and hence it is necessary to erect a grid and record X , Y , Z , T , F only at intersections. The spacing of the points in this array is fairly critical. Geological data are distributed throughout the volume, and hence the closer the spacing of points, the more detail that can be recorded. However the amount of data to be handled increases rapidly as the point of spacing is reduced, and so an optimum spacing of points must be taken to balance these two factors. Data thus recorded are in a form suitable for storage and retrieval procedures. The general scheme should thus be capable of development for computer processing.

A variety of data plots can be extracted from the data when they are stored in this form. The most useful for visual presentation are plots which show the variation of one or more parameters on a two

dimensional reference frame based on two of the other parameters. Ten possible pairs of reference axes can be selected from the five parameters X, Y, Z, T, F. These are listed in Table 1. Only certain of these possible plots are of real value (indicated in Table 1 by **), and these are further outlined in Table 2, which is self-explanatory. Other plots could be constructed within this general framework as required. Tables 1 and 2 together thus illustrate how, within the five-parameter data recording scheme developed, the various familiar data plots are inter-related. They also indicate another type of plot, involving time as one axis. These, particularly T-X plots, are of interest in tectonic analysis.

There is no theoretical reason why three- and four-dimensional plots cannot be constructed. A four-dimensional plot with X, Y, Z, T axes could record all the geological history of an area. In such a plot a time surface passing along the time axis would move through the X-Y-Z cube in such a way that it would simulate the geological history of the region in terms of F-variation in space and time. Such hypothetical multi-dimensional plots are difficult to visualise, but computer processing might be usefully applied to them.

I have developed the above scheme in relation to a hypothetical continental margin, in order to illustrate how various geological data plots are inter-related. The general principles are however applicable to any situation. In more complex areas which have undergone deformation similar recording techniques could be developed, but in such terrains a particular unit of geological information formed at a point at a certain time will not remain fixed with respect to the X-Y-Z frame.

TABLE A.1: AXIS PAIRS FOR TWO-DIMENSIONAL DATA-PLOTS

<u>Axes</u>	<u>Representation</u>
(X - Y)**	areal distribution (z,T,F)
(X - Z)**	vertical cross-section (T,F)
(X - T)**	time-space cross-section (F)
(X - F)	F-factor variation along X
(Y - Z)	vertical section along strike of margin
(Y - T)	time-space strike section
(Y - F)	F-factor variation along Y
(Z - T)**	time-space vertical column
(Z - F)	F-factor variation in vertical column
(T - F)	F-factor/time variation

**Useful plots when combined with any one of parameters in brackets (see Table A.2)

TABLE A.2: EXAMPLES OF TWO-DIMENSIONAL DATA-PLOTS

<u>Axes</u>	<u>Variable</u>	<u>Specified</u>	<u>Interpretation</u>
X - Y	$F(fm,r)$	at T_0	present-day geological map
	$F(fm,r)$	at T_1	paleogeological map at T_1
	$F(r)$	at z_0	present shoreline
	z	at T_0	bathymetric and topographic map
	z	at T_1	structure contour map of time-stratigraphic surface t_1
	z	interval T_1-T_2	isopach map of interval t_1-t_2
	z	on $F(s)$	structure contour map on rock-stratigraphic surface s
	z	between s_1 and s_2	isopach map of formation between s_1 and s_2
	T	on $F(s)$	time-variation on rock-stratigraphic surface s
X - Z	$F(fm,r)$	at T_0	present-day cross-section with rock-stratigraphic boundaries
	$F(fm,r)$	at T_1	reconstructed cross-section at T_1 , with rock-stratigraphic boundaries
	$F(fm,r),t$	at T_0 or T_1	as above, with superimposed time-lines - indicates diachronous deposition

(NOTE: for each X - Z plot, the Y - co-ordinate must be specified - by progressively increasing Y, a series of sections can be obtained)

TABLE A.2 (continued)

X - T F(fm,r) at T_0 time-space plot which emphasises tectonic evolution

(NOTE: for each X - T plot, the Y - co-ordinate must be specified)

Z - T F(fm,r) x,y, T_0 vertical facies variation and rate of sediment accumulation

(*SPECIAL CASE*: plot F(fm,r) and T along one axis Z - section in vertical borehole)

Y - Z Similar sections to X - Z, but taken along "strike" at specified X - indicate diachronous deposition of lithostratigraphic units along margin, etc.

Y - T Time-space plot along margin - indicates diachronous tectonic developments along "strike", etc.

Thus the distortion of the X-Y-Z frame back through time would have to be deduced, which would be difficult except on a very general level in practice. The scheme does however provide a theoretical basis for the analysis and storage of data, and shows how various plots can be extracted from it. The most interesting of these with respect to tectonic analysis are the time-space plots, which I have developed as very useful tools in tectonic interpretation.

A.3: TIME-SPACE PLOTS

The basic concept of a time-space plot is embodied in several diagrams used in stratigraphy. The most common is the geological correlation chart. This has a vertical time axis, and usually stratigraphic columns from different localities are placed side by side along the horizontal axis. Each column is generally a sedimentary succession, correlated with the others, and the diagram shows unconformities, facies variations etc. as appropriate. These are not true time-space diagrams in that the "space" axis is not a continuous section. Another commonly used diagram shows diachronous facies variation along a cross section by using time rather than depth as the vertical axis. These have the same properties as the time-space plots developed here, but are generally restricted in their use to narrow time ranges and undeformed geological sections.

In recent papers dealing with the geological applications of plate tectonics (Dewey, 1969b; Dewey & Bird, 1970a) more elaborate time-space plots have been constructed to illustrate plate tectonic models for geosynclines and orogenic belts. These have sought to

show how sedimentary, structural, igneous and metamorphic events are inter-related in time and space. In their analysis of the Appalachian orogen, Bird & Dewey (1970) use a similar time-space diagram to illustrate their proposed plate tectonic model for the evolution of this orogen, incorporating names of rock units, thrust sheets, etc., thus relating the model directly to the actual geology of the area. At the present time, this appears to be the stage that the use of time-space plots has reached in the literature.

A.3.1: *CONSTRUCTION OF TIME-SPACE PLOTS*

Time-space plots are members of a family of geological data plots. In the previous section I have shown theoretically how an objective description can be made, based on the recording of the age (T) and rock-factor (F) at each of a network of points in a three-dimensional array, and thus a variety of plots can be quantitatively extracted from it by specifying a pair of axes (from X, Y, Z, T, F) and then plotting the variation of other selected parameters in this framework. There is thus nothing inherent in the concept of time-space plots such that they cannot be constructed from the normal spectrum of geological information gathered in an area. In particular, they do not depend upon the validity of any tectonic models. They are simply another, hitherto little used, method of presenting geological data.

The basis of a time-space plot is a perpendicular pair of axes. The vertical axis represents time and the horizontal axis "space" (i.e: a section across the area under consideration). Any horizontal line across the plot thus represents a single instant in

time. It is most convenient in practice to have time decreasing *upwards*, thus retaining geological events in order of superposition as seen in real sections. The time axis is divided off into geological periods, and subdivided into epochs or smaller units if required, to cover the time range of the area. Absolute ages can be used to give a linear scale, based on an absolute/relative age correlation chart (e.g: Geological Society Phanerozoic time-scale, 1964). Time lines (e.g: period boundaries) are ruled across as horizontal lines, and the resulting framework is the basis for plotting the geological data.

In order to plot data from a particular area, it is necessary to select a line of section for the "space" axis. The most useful property of a time-space plot is that it shows the evolution of a *geological* cross section through time. A time-space plot can be compiled for any chosen section, but with a view to later use in tectonic synthesis, a section containing the maximum sedimentary and structural information should be selected. For essentially linear geological features (continental margins, orogenic belts) such a section will in general be normal to the linear trend.

A.3.2: APPLICATION TO ESSENTIALLY UNDEFORMED SEDIMENTARY ACCUMULATIONS

In terms of the scheme developed above for objectively recording the geology of an aseismic continental margin, the most useful section for a time-space analysis is one normal to the "strike" of the margin, that is from craton to ocean floor (a T-X section). If the assumption that continental margins form by rifting of a continental craton and ocean basin formation is valid, then such a

section is perpendicular to the rift axis, and should therefore record the maximum of information relevant to pull-apart and continental margin evolution. In contrast, a section parallel to the rift axis (a T-Y section) records mainly diachronous events along the length of the margin, which are in general much more gradual. Sections of intra-cratonic, undeformed sedimentary basins should similarly be selected to cut across basin-forming structures as far as possible.

A conventional section (with depth as the vertical axis) of such a continental margin or sedimentary basin generally shows mainly stratigraphic and structural data - i.e: details of lithology and age at each point, faults, etc. Conversion of such a section to a time-space section involves the identification of time lines through the section, and then "straightening" these out. Information regarding sediment type will shift vertically to lie on its appropriate time line on the plot, providing there has been no overall change in length of the section during the time period analysed. If there has (e.g: by normal faulting during rifting) the length of the section will change with time on the time-space plot. These general principles are illustrated diagrammatically in Fig. A.4. One of the major disadvantages of two-dimensional time-space plots is also apparent in Fig. A.4 - it is impossible to show the *thickness* (and thus record the rate of deposition) of sediments between time lines. Theoretically this can be overcome by using a three-dimensional plot with the third axis "rate of deposition" (and erosion in the negative sense), but such plots would be more

difficult to handle and visualise and will not be discussed here.

The method of representation of data must also be established before plotting it. Lithological information can be illustrated by using conventional symbols for sandstones, shales, limestones, etc., with gradational facies variation indicated by intermingling. Unconformities may be shown by wavy lines and periods of non-deposition by vertical ruling. Examples of these and other of the symbols which I have adopted in my own plots are shown in Fig. A.5. It is stressed that any time-space diagram must have a comprehensive key, indicating the adopted symbolism, to be of maximum communicative value.

Structural features can also be portrayed. Faults are shown as heavy lines over their period of activity; episodes of folding as broad wavy traces, etc. (Fig. A.5). If faults or folds extend or shorten the section, the apparent length of the section on the space axis will correspondingly change (Fig. A.4). Other features such as direction of currents (sediment provenance), fossil occurrences, igneous intrusive and extrusive rocks, etc., can be plotted if required (Fig. A.5). It should be noted that *extrusive* igneous rocks will occur in their correct relative position on the plot, immediately above the sediments on which they lie, whereas for *intrusive* rocks the plot shows only their time of intrusion and gives no indication as to what position in the sedimentary column they now occupy.

A.3.3: APPLICATIONS TO GEOSYNCLINES AND OROGENIC BELTS.

The fore-going outline of how time-space plots are derived has related largely to their application to present day aseismic continental margins and sedimentary basins, but also illustrates the theoretical basis of the plots. The same general principles can be used to construct time-space plots of more complex areas such as orogenic belts, and by passing back in time, to the "geosynclines" from which they have formed. Again, with a view to later tectonic interpretation, a section should be chosen so as to include the greatest amount of relevant information. For an orogenic belt (or geosyncline), essentially long narrow features, a section directly across the belt is potentially most useful. A section along the length of the belt will record mainly diachronous deformation along the orogen.

In orogenic belts, the amount of relative horizontal movement of units across the belt is generally large, and may not even be accurately determinable in many cases. Very detailed work may allow estimates of the shortening involved to be derived, but often only *relative* movements and position of points can be determined. In these cases the time-space plot will not be so quantitative, but still allows the time relations of events and their relative positions in space to be accurately displayed. Bulk transport of rocks as nappes, thrusts and gravity slides must also be taken into account.

In the early geosynclinal stages, most of the data will be of the type already discussed, but in orogenic belts the variety of

geological data is greatly increased. Lithological and igneous data can be portrayed as previously indicated (Fig. A.5). In addition, data pertinent to structural deformation and metamorphism can be plotted. Again it is necessary to first establish a suitable symbolism to display this information. Stages of deformation, as worked out by structural analysis, can be plotted in time and over the length of section they affect. Details of structural styles etc. cannot be adequately plotted, but such data could be included in an accompanying table. Metamorphic events can similarly be plotted with respect to time and the area they affect. It is not possible to show the grade of metamorphism, as this is in part a function of depth, but by using suitable symbols it should be possible to indicate the metamorphic facies (i.e: the variation of geothermal gradient with time). Examples of some of these symbols used in my plots are also shown in Fig. A.5.

In deformed belts it may not be possible to have a quantitative space dimension. This will generally become even less possible as the complexity of the belt increases. However it is still possible to plot the *relative* positions of events in space, which is the most important point. All the data which is plotted is the type of "conventional" information which geologists always seek in orogenic belts - i.e: the relation of events in space and time. Thus once again it is stressed that time-space plots only use the established spectrum of geological data, and there is no reliance on any tectonic model. The time-space plot merely records

data, and portrays it in a form very suitable for later tectonic synthesis.

A.3.4: *CONSTRUCTING ACTUAL TIME-SPACE PLOTS.*

Throughout the discussion so far I have assumed that data is being collected from a single-plane section. However in reality this is virtually impossible, as the field geologist studies an *area*, and works out the rock relationships, structural history, etc. by considering data from this area. The information from which a time-space plot is derived will therefore generally be in the form of maps, diagrams and literature, which together contain the essentials of the geology of the area to the extent that it is known.

This information does not all lie on one plane, but nevertheless can be used to construct a single time-space plot covering the area. In this case the final diagram is really a "stack" of individual time-space plots of actual sections, such that bits of data from each are condensed onto a single plot. For the major features such as continental margins and orogenic belts, this will not in fact cause much distortion of data, as these features are essentially linear anyway, and the sections can be selected normal to their linear trend. If any events are severely diachronous along this trend, they will tend to "spread" in time on the two-dimensional plot.

It is possible to conceptualise further development of time-space plotting by moving into the realms of multi-dimensional

plots. Thus for example three axes of time, length and breadth could be used, thus illustrating the variation of data with time in two dimensions. A four-dimensional plot (T, X, Y, Z) would also allow data relevant to depth, and hence rates of deposition, metamorphic grades, etc., to be plotted with respect to time. All these diagrams have the same property - that a time line, plane or volume can be selected which shows the stage of development of the geology of an area at a particular instant. Passage of such a line, plane or volume through time then simulates the geological evolution of a region.

The process of "stacking" data from different cross sections has considerable advantages in practical terms. A single section records only data at the present erosional surface, and perhaps a little sub-surface data if available (a considerable amount of sub-surface data may be obtained by methods such as seismic profiling in certain cases). However, passing along an orogenic belt, the present erosional surface does not necessarily correspond to any previous time surface, but usually transects many older time surfaces, thus revealing much of the history of the area through time. This basic principle has been used with particular success in the analysis of New Zealand, where there is a fairly complete progression from north to south through past time planes covering the whole evolution of New Zealand (see Chapter 3.2.5.).

A word regarding the practicalities of constructing time-space plots. I have found that the best method is to use a large blackboard

and coloured chalks. The time grid is first drawn, and then data can be plotted on the board. This method allows great flexibility in planning the proportions and layout of the plot. Once a rough outline of the major elements has been made, this can be redrawn on a second blackboard and then the additional detailed data plotted. If modifications to this outline are required as a result of plotting the details, these can easily be effected by erasing and redrawing the relevant parts. The final version can be photographed, printed at a convenient size, and then a line copy prepared by tracing from the photograph.

A.4: METHODS OF TECTONIC SYNTHESIS USING TIME-SPACE PLOTS.

A geological study of a region ultimately reaches the stage where a tectonic synthesis is made. In other words, all the assembled geological data is now synthesised, in terms of a particular tectonic hypothesis, to give an evolutionary model for the region. There have been many attempts at tectonic synthesis, and most of these have reflected the tectonic hypothesis favoured at the time they were made. The rapid development of plate tectonics over the last few years has probably produced more "synthesisers" than any other theory.

Commonly, a tectonic synthesis is produced more by intuition than method. For example, the subduction zones of plate tectonics have particular features associated with them (calc-alkaline volcanics, ophiolites, etc.). Study of the geology of an old

orogenic belt might reveal similar assemblages, and hence suggest the locations of "fossil" subduction zones. Other geological features (structure, metamorphism, etc.) might then be interpreted in terms of the subduction model, and so a synthesis is constructed. This is the most usual approach, and can produce meaningful results. However, in the foregoing sections I have tried to look objectively at the whole process leading to tectonic synthesis, to see if there is a methodological approach which can be used. This appears to be the case, as I shall show.

Many tectonic syntheses will continue to be produced by the "intuitive" approach - in other words entering the flow-scheme at some arbitrary point. This may well be the initial breakthrough in the synthesis of a region, but I would stress the value of outlining the theoretical approach which should be followed, in order to make the basis of the synthesis quite clear. If a synthesis is constructed intuitively in the first instance, it should then always be critically evaluated in terms of the theoretical scheme I have outlined (Section A.1), as a means of testing its validity. Use of time-space plots, however, appears to present a new route to tectonic synthesis which may to a large extent allow a methodological approach to be used in the first instance.

In terms of the simple flow-scheme outlined (Fig. A.1), the purpose of tectonic analysis is to present geological data in a form suitable for synthesis. The end result of the analysis should be presentation of this data in the most objective form possible. Time-space plots have this desired property, in that they are

derived straight from the data, and are a very compact and useful means of portraying a great variety of data on a single diagram. Time-space plots can also be constructed for the theoretical models based on tectonic hypotheses. I have mentioned examples of such plots prepared by Dewey (1969b) and Dewey and Bird (1970a) based on plate tectonic models. Hence, whereas tectonic synthesis might proceed by a simple scheme (e.g: Fig. A.1), it is now possible to devise a more elaborate scheme in which time-space plots are used as an intermediate step (Fig. A.6).

The basis of this scheme is the *comparison* of actual and theoretical time-space plots. These can be two-, three-, or four-dimensional plots. The results of tectonic analysis are presented as an actual time-space plot. Some data can be directly plotted without prior interpretation, thus increasing the objectivity of the analysis (see Fig. A.6). A tectonic hypothesis is then adopted, and an *actualistic model* for the particular area is made. Inductive reasoning can be used in constructing both the hypothesis and the model, as it will later be critically evaluated against the basic data. A time-space plot of the model is then made. The two are now compared with each other (Fig. A.6). If they are identical, then the model simulates the geological data, and can be adopted as a tectonic synthesis of the data. Further evaluation of the synthesis is possible by using feed-back and predictive tests as before. If the plots are different, *only* the model plot can be changed, so the model must be modified and a new time-space

plot constructed. This loop can continue until the real and model plots are the same. If this proves impossible, then it suggests that no actualistic model can be made on the basis of the hypothesis, and therefore the latter may be wrong and must be rejected.

Use of time-space plots in this way has several advantages: basic geological data is objectively recorded, and once this is so it cannot be "changed" to fit the model; there is no possibility of using the model to interpret basic data as in Figs. A.2, A.3; use of "geosynclinal" nomenclature, and its inherent problems, is now an unnecessary step; a tectonic hypothesis can itself be evaluated by testing its ability to synthesise geological data. Time-space plots can therefore be used in tectonic analysis as a very useful technique, which I have used in several different situations in the south-west Pacific region.

A.5: GENERALISED EXAMPLES OF TECTONIC SYNTHESIS USING TIME-SPACE PLOTS

Finally in this chapter I will briefly outline two applications of this scheme of tectonic analysis and synthesis which could be used in real situations. These relate to the synthesis of a continental margin and an orogenic belt, in terms of the continental drift and plate tectonic hypotheses. In planning any such synthesis using the scheme, the first step is to prepare a *flow chart* showing all the stages, so that the various operations to be carried out, and the whole approach, are made quite clear. These flow charts may become large and complex, but should still retain the basic

elements of the simple scheme (Fig. A.1). When using time-space plots, it must also be remembered that only two-dimensional ones will be easy to use, and hence several may be necessary to display all the data and illustrate the actualistic model. Comparison of plots then implies that all of these must be used.

A.5 1: *CONTINENTAL MARGINS*

Figure A.7 represents a flow-chart for tectonic synthesis of a continental margin. The situation envisaged is an aseismic, "Atlantic-type" continental margin - i.e: a margin at which sediments, derived largely from erosion of a continental area, are accumulating out onto oceanic crust. Geological data (Phase 1) is of three types - geophysical (largely relating to structural configurations); borehole logs (providing stratigraphic control); direct mapping of sub-aerial exposures onshore. Tectonic analysis (Phases I - III) then leads to a geological history, which can be portrayed as maps, sections and accompanying narrative. All of this information is then used to construct time-space plots (see Section A.3.2).

From inductive geological reasoning, the hypothesis of continental drift has been proposed, and it is supposed that a single continent can split apart, initially forming a rift valley and ultimately a new ocean basin. An actualistic model is prepared on the basis of this hypothesis, and derived time-space plots constructed. The two sets of time-space plots are now compared. If they are identical, or have an acceptable and pre-specified similarity, a tectonic synthesis of the continental margin based on

the continental drift hypothesis can be made. This is then evaluated against the original data using feed-back and prediction tests, to see if it is really consistent with all the data. If it is not acceptably similar, the model parameters are varied, new time-space plots derived, and new comparisons made. This process continues until either an acceptable level of similarity is obtained, or it is not possible to further alter the model to achieve this. In the latter case, the original hypothesis must either be modified, or rejected and a new hypothesis advanced.

A.5.2: *OROGENIC BELTS*

Figure A.8 is an example of a flow chart leading to tectonic synthesis of an orogenic belt. An orogenic belt is a zone in which orogeny, or mountain-building, has taken place. Original sedimentary accumulations (usually referred to as geosynclines in this context) have undergone deformation due to extensive faulting and folding, accompanied by regional metamorphism and igneous activity. Data collection in such terrains generally involves field-mapping, collecting of specimens and laboratory investigations (thin section study, chemical analysis, structural analysis, etc.). As a result (Phase II) it is possible to ascertain certain structural, metamorphic and igneous events, and to elucidate something of the types and stratigraphic relationships of the original sedimentary rocks. Integration of these results leads to a geological history (Phase III). Time-space plots can then be constructed from the data (see Section A.3.3).

In this case a plate tectonic hypothesis of orogenesis related to subduction zones has been proposed, an actualistic model constructed, and time-space plots derived from it. Comparison of the two sets of time-space plots is then carried out as in the previous section, until either an acceptable synthesis of the orogenic belt is obtained, or the plate tectonic hypothesis is modified or rejected entirely. Feed-back and predictive tests are applied as before to evaluate the synthesis. In particular, prediction of new data which can be sought can be very useful, and can point to the need for information about specific problems as a guide to planning further work in such complex regions as orogenic belts.

Although both the above flow charts are complicated, careful inspection will show that they still obey the basic rules discussed in Section A.1. The use of time-space plots has allowed total discrimination between hypothesis and analysis. It is only after a model has been erected to attempt to simulate the results of the tectonic analysis that the two lines are brought into contact by comparing time-space plots, and even then further evaluation is possible using various tests. The scheme also allows any tectonic hypothesis to be evaluated against geological data, and it should thus be useful in testing the validity of the plate tectonic hypothesis.

I would hope that further use of the methods which I have outlined will prove a useful new approach in tectonics, and allow some order to be brought into a previously confusing and very subjective part of geology.

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2. UNPUBLISHED MATERIAL AVAILABLE

UNDER PETROLEUM SEARCH SUBSIDY ACT

(See pages 4.2, 4.3)

(a) AEROMAGNETIC SURVEY (AMS)

Bass Strait and Encounter Bay. Haematite Explorations Pty. Ltd., 1960-1.
[62/1710 and 62/1711]. P.S.S.A. Pub. 60 (1965).

(b) SEISMIC SURVEYS (SS)

Bass Area (EB)SS Esso Exploration Aust. Inc., 1965 [65/4588].

Bass (ED-67) SS Esso Exploration Aust. Inc., 1967 [67/11196].

Bass (EF-68) SS Esso Exploration Aust. Inc., 1968 [68/3014].

Bass (B69A) SS Esso Exploration and Production Aust. Inc., 1969

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Eastern Bass Strait (ES and ET) SS Esso Exploration and Production Aust.
Inc., 1966 [66/11070].

East Gippsland/Tasmania/Bass Strait SS Magellan Petroleum Australia Ltd.,
1969 [69/3023].

East Tasmania T69B SS Esso Exploration and Production Aust. Inc., 1969
[69/3001].

Gippsland (EG) SS Esso Exploration Aust. Inc., 1965. [65/11045].

Gippsland (EC-67) SS Esso Exploration and Production Aust. Inc., 1968.
[67/11184].

Gippsland (EH-68) SS Esso Exploration and Production Aust. Inc., 1968.

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Gippsland (G69A) SS Esso Exploration and Production Aust. Inc., 1969.

[68/3058].

Hawkesdale SS Shell Development (Aust.) Pty. Ltd., 1969 [68/3053].

King Island East (EK-65) SS Esso Exploration Aust. Inc., 1966. [65/11037].

Otway Marine (EO) SS Esso Exploration and Production Aust. Inc., 1967

[65/11052].

Otway (EP-67) SS Esso Exploration and Production Aust. Inc., 1967/8

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Otway (ER-68) SS Esso Exploration and Production Aust. Inc., 1968

[68/3036].

Otway (EU-68) SS Esso Exploration and Production Aust. Inc., 1968

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Otway (O69A) SS Esso Exploration and Production Aust. Inc., 1969

[69/3003]

Otway (O69B) SS Esso Exploration and Production Aust. Inc., 1969

[69/3061].

West Tasmania T69A SS Esso Exploration and Production Aust. Inc., 1969

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(c) WELL COMPLETION REPORTS (WCR)

Anglesea WCR Oil Development N.L., 1962 [62/1217].

Argonaut WCR. Esso Exploration and Production Aust. Inc., 1968 [68/2018]

Barracouta WCR. Esso Exploration Australia Inc., 1965 [

P.S.S.A. Pub. 76

Bass-1 WCR. Esso Exploration Australia Inc., 1965 [65/4167] P.S.S.A.

Pub. 83

Bass-2 WCR. Esso Exploration Australia Inc., 1966 [66/4187] P.S.S.A.

Pub. 83

Bass-3 WCR. Esso Exploration Australia Inc., 1967 [67/4241].

Beachport WCR. South East Oil Syndicate Ltd., 1961 [62/1211].

Casterton No. 1 WCR. Planet Exploration Co. Pty. Ltd., 1965 [65/4135].

Caroline WCR. Alliance Oil Development Aust., 1967 [66/4222].

Crayfish WCR. Esso Exploration Australia Inc., 1967 [67/4266].

Duck Bay WCR. Woodside (Lakes Entrance) Oil Company N.L. and Arco Limited,
1964 [64/4014] P.S.S.A. Pub. 77

Fergusons Hill WCR. Frome-Broken Hill Co. Pty. Ltd., 1964 [63/1061].

Flaxmans WCR. Frome-Broken Hill Co. Pty. Ltd., 1961 [62/1074] P.S.S.A.
Pub. 62.

Garvoc WCR. Interstate Oil Ltd., 1968 [68/2020].

Geltwood Beach WCR. Beach Petroleum N.L., 1963 [63/1323]. P.S.S.A.
Pub. 65.

Hawkesdale WCR. Shell Development (Aust.) Pty. Ltd., 1969 [69/2032].

Heathfield WCR. Planet Exploration Co. Pty. Ltd., 1964 [64/4019].

Lake Eliza WCR. Esso Exploration and Production Aust. Inc., [69/2027].

Lake George WCR. Esso Exploration and Production Aust. Inc., 1969 [69/2030].

Lucindale WCR. Esso Exploration and Production Aust. Inc., 1969 [69/2034].

Kalangadoo WCR. Alliance Oil Development Aust. N.L., 1965 [65/4153].

Mount Salt WCR. Oil Development N.L., 1962 [62/1100] P.S.S.A. Pub. 64

Moyne Falls WCR. Shell Development (Australia) Pty. Ltd., 1969 [69/2032].

Nautilus WCR. Esso Exploration and Production (Australia) Inc., 1968
[68/2008].

Nerita WCR. Shell Development (Aust.) Pty. Ltd., 1967 [67/4258] P.S.S.A.
Pub. 88.

Pecten WCR. Shell Development (Aust.) Pty. Ltd., 1967 [67/4239] P.S.S.A.
Pub. 88.

Penola WCR. Oil Development N.L., 1961 [62/1071] P.S.S.A. Pub. 42.

Prawn WCR. Esso Exploration Aust. Inc., 1968 [67/4273].

Pretty Hill WCR. Frome-Broken Hill Co. Pty. Ltd., 1962 [62/1115].

Purrumbete WCR. Interstate Oil Ltd., 1968 [68/2021].

Robertson -1 WCR. Alliance Oil Development Aust., 1967 [66/4237].

Robertson -2 WCR. Alliance Oil Development Aust., 1967 [67/4250].

Sherbrook WCR. Frome-Broken Hill Co. Pty. Ltd., 1963 [63/1045].

South-west Bairnsdale WCR. Woodside (Lakes Entrance) Oil Company N.L.
and Arco Limited, 1963 [62/1224] P.S.S.A. Pub. 77.

Voluta WCR. Shell Development (Aust.) Pty. Ltd., 1967 [67/4263].

Wellington Park WCR. Woodside (Lakes Entrance) Oil Company N.L., 1962
[62/1077] P.S.S.A. Pub. 71.

Woolsthorpe WCR. Interstate Oil Ltd., 1968 [68/2019].